

WESTERN PENNINES

Field Guide

Edited by

W. A. MITCHELL

Quaternary Research Association



1991

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Cover Illustration: Geomorphological map showing drumlins and meltwater channels in Grisedale and around Garsdale Head, Yorkshire (from Mitchell, 1991).

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ISSN 0261-3611

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Word Processing and formatting : W. A. Mitchell, Luton College of Higher Education.

Printing : John Winn, M1 Press Ltd., Trent Business Centre, Canal Street, Long Eaton, Nottingham, NG10 4HQ.

Recommended reference : Mitchell, W. A. (editor), 1991. *Western Pennines: Field Guide*. Quaternary Research Association, London.

QUATERNARY RESEARCH ASSOCIATION

WESTERN PENNINES : FIELD GUIDE

Edited by W. A. Mitchell

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ACKNOWLEDGEMENTS

The editor would like to thank Jim Rose for his considerable help, advice and encouragement in the compilation of this guide book.

The logistic support of the Faculty of Applied Sciences, Luton College of Higher Education in the preparation of this field guide is gratefully recorded; in particular thanks are extended to Tessa Smart and Jackie Errington for typing the tables, and to Donna Bakewell for drawing some of the diagrams and general cartographic advice. Thanks are also due to Jenny Matthews for looking after the references with her usual efficiency.

The contributors are grateful to the various landowners for access to the land to visit the sites.

Jim Rose would like to thank the many students from Birkbeck College and Royal Hooloway and Bedford New College, University of London for collecting the data recorded at Widdale Side and providing stimulating discussion about its significance.

Research in the area has been supported by grants from the Central Research Fund, University of London and Luton College of Higher Education.

Thanks are also given to the Geological Society and the Yorkshire Geological Society for permission to publish Figures 2 and 3 and Table 11.

Ordnance Survey maps covering the Western Pennines include the 1:50,000 Landranger Series of Great Britain, Sheets 98 (Wensleydale and Wharfedale) and 91 (Appleby).

CONTENTS

Part 1: Background Information

Quaternary of the Western Pennines: a review and physical background	<i>W.A. Mitchell and J. Rose</i>	1
---	----------------------------------	---

Geomorphological Mapping	<i>W.A. Mitchell</i>	19
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Dimlington Stadial Ice Sheet in the Western Pennines	<i>W.A. Mitchell</i>	25
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Loch Lomond Stadial Glacial Landforms and Palaeoglaciological Reconstruction	<i>W.A. Mitchell</i>	43
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Part 2: Site Descriptions

Widdale	<i>W.A. Mitchell</i>	55
Widdale Side	<i>J. Rose</i>	61
Swarth Fell	<i>W.A. Mitchell</i>	66
Wild Boar Fell	<i>W.A. Mitchell</i>	71

Grisedale	<i>W.A. Mitchell</i>	79
Loven Scars and Hangingstone Scar	<i>W.A. Mitchell</i>	82

Bluecaster	<i>W.A. Mitchell & T.P. Buggie</i>	91
Cautley Crags	<i>W.A. Mitchell</i>	94
Garths	<i>A.R. Gunson</i>	99
Combe Scar	<i>A. R. Gunson & W.A. Mitchell</i>	104

PREFACE

Northern England has been the location for three previous QRA short field meetings in the last ten years. In 1981, the QRA met at Appleby to consider recent work which had been completed in the eastern Lake District and southern Vale of Eden. In 1985 a QRA/IGU meeting was organised to visit the periglacial landforms on the Alston Block and the Lake District and the following year there was the first BGRG/QRA field meeting to discuss river landforms and sediments in the Northern Pennines.

Surprisingly with all this activity, there are still large areas of the Pennines for which very little is known regarding their Quaternary evolution and the sites visited on previous trips have proved difficult to place within a regional context. Much of the regional history is, therefore, still based on ideas proposed in the latter part of the last century and in the early part of this century.

This trip has been organised to demonstrate some of the recent work which has been completed in a classic area of the Pennines - the western part of the Yorkshire Dales and which, coincidentally, is traversed by the Settle to Carlisle Railway from Blea Moor Tunnel to Ais Gill Summit.

The field guide has been designed to provide background information on a number of the key themes which will be developed during the field meeting (Part 1) and by describing critical sites for the understanding of glaciation and its impact on this upland area (Part 2).

Jim Rose and Wishart Mitchell, August, 1991

QUATERNARY OF THE WESTERN PENNINES : A REVIEW AND PHYSICAL BACKGROUND

W.A. Mitchell and J. Rose

The Western Pennines are defined in this guide as the northwestern part of the Yorkshire Dales, particularly upper Wensleydale, Garsdale, Dentdale and the adjoining upland areas (Fig. 1). The area is roughly comparable to the geological structural unit of the Askrigg Block, which is delimited by the Stainmore Disturbance, Dent Faults and Craven Fault systems (Fig. 2), although sites will be visited in the eastern Howgill Fells and Rawthey valley which lie on the western side of the Dent Fault. This is a much more restricted areal definition of the Western Pennines than is normally employed in the literature (*cf* Johnson, 1985) but reflects an area which has been the subject of a recently completed research project (Mitchell, 1991). Discussion will also focus on the history of glacial events in the southern part of the Vale of Eden (Letzer, 1978, 1981, 1987) to place the Western Pennines into a regional perspective. However, an overview of glacial events in northern England is some way in the future since little detailed fieldwork has been completed for many other critical areas.

The Quaternary of the Western Pennines is important from a number of different points of view. Firstly, it is a critical region with respect to ice dispersal patterns in northern England. Recent models of the last (Dimlington Stadial) British ice sheet are still at a preliminary stage and because of the inadequate field information available, published maps show conflicting interpretations about the existence of regional ice centres and their role in the dynamics of the last British ice sheet. For instance, no regional ice centre is recorded for northern England in Boulton, *et al.*, (1977) but is generally identified over the Lake District in the maps published in a later work (Boulton, *et al.*, 1985). Secondly, apart from pollen analysis at limited sites (Walker, 1955a, 1955b; Gunson, 1966) and the abundance of work on environmental change in Britain following the wastage of the Late Devensian (Dimlington Stadial) ice sheet (*cf* Lowe and Walker, 1984), the fact remains that the Western Pennines is one of the few upland areas which has yet to be studied systematically to produce a coherent interpretation of environmental conditions during the Lateglacial period. It is also one of the last upland areas in the United Kingdom, if

not Ireland (Gray and Coxon, 1991), in which the Loch Lomond Stadial glaciers have been mapped in detail (Mitchell, 1991).

Finally, there is no reference in published work to the numerous mass movement features that have been found throughout the Western Pennines (Mitchell, 1991), although large scale slope failures have been noted in upper Swaledale (Rose, 1980) and in more southerly areas of the Pennines (Johnson, 1987). Within the area of the Western Pennines, such landforms range from very large deep seated rock slope failures, through mudslides to smaller slope failures in superficial deposits. Information on such features is important in an assessment of the impact of glaciation on an area as well as being critical to the overall understanding of paraglacial landscape evolution since the disappearance of the glaciers.

Relief

Wensleydale and Garsdale form the central west-east axis of the area and are joined at Garsdale Head by Mallerstang which is a north-south trending through valley with the interfluves forming distinctive Pennine fells (Fig. 1). The highest ground within the area is formed by Great Shunner Fell (716m), to the north of Wensleydale and Whernside (736m) to the south of Dentdale (Fig. 1). The most extensive upland area is Baugh Fell (676m), which covers 36 sq. km between the River Rawthey and River Clough. Widdale Fell (672m) forms a major upland between Widdale, Dentdale, upper Garsdale and Wensleydale, but most of the other watershed areas are less extensive. The northern interfluve of Wensleydale is a linear stretch of moorland between Hugh Seat (688m) on the eastern side of Mallerstang to the summit of Great Shunner Fell. The southern watershed of Wensleydale is also marked by a narrow stretch of plateau from Grove Head to Fleet Moss with a highest point at Dodd Fell (668m).

Around Sedbergh, the altitude of the floors of Garsdale and Dentdale are just over 100m and rise eastwards to 300m at the head of each dale around Garsdale Head and Dent Head respectively (Fig. 1). In Wensleydale, the valley floor has a height of c. 200m at Hawes with a westward rise to Garsdale Head at 313m where it is continued northwards as an area of low ground which rises to 364m at Ais Gill Summit. In general, the valleys are often incised between 300 and 400m below the general level of the mountain summits.

The main rivers in the Western Pennines include the River Ure which flows east

through Wensleydale with a source on Lunds Fell close to the source of the River Eden which drains north through Mallerstang and the Vale of Eden (Fig. 1). Of the other rivers within the area, both the River Clough and River Rawthey rise on Baugh Fell with the former flowing south east down Grisedale, and then southwestwards through Garsdale, whilst the River Rawthey flows northwest, and then south where it joins the River Clough just east of the town of Sedbergh. To the west of Sedbergh, the River Rawthey is joined by the River Dee, which drains Dentdale, before they join the River Lune (Fig. 1).

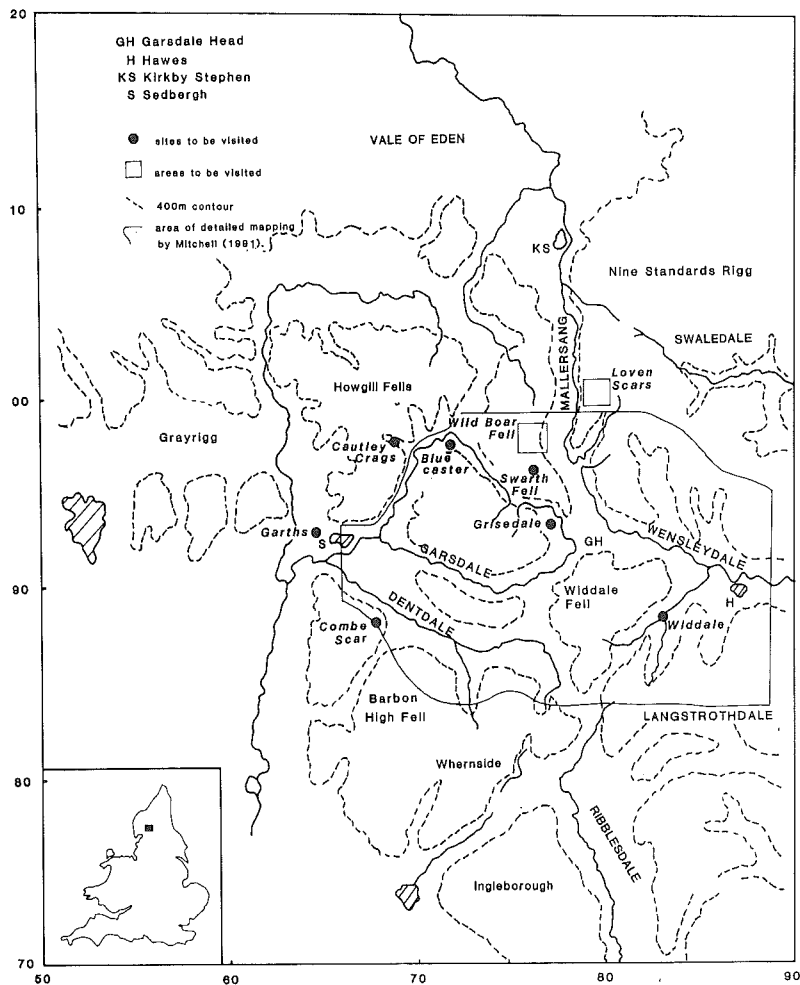


Figure 1. Topography and drainage of the Western Pennines showing site locations and areas to be visited. Scale and orientation given by National Grid co-ordinates.

Pre-Quaternary Geology

The geology of the Western Pennines reflect the two major lithological provinces first identified by Sedgwick (1835b), with the Ordovician/Silurian rocks of the Lake District being separated from the Carboniferous rocks of the Pennines by the Dent Fault System (Fig. 2) (Underhill, *et al.*, 1988).

The main sequence of pre-Carboniferous rocks exposed in the area of the eastern Howgill Fells (Fig. 2) is dominated by the Stockdale Shales and Brathay Flags, a thick sequence of laminated graptolitic mudstones with calcareous nodules (Rickards, 1978). Overlying these in the Howgill Fells are sandstones, termed the Coniston Grits (Fig. 3) (Rickards, 1978). All of these rocks were formed at various stages during the Silurian Period and show deformation and low grade metamorphism associated with Caledonide events (Soper and Moseley, 1978). An extensive granite batholith underlies the Lake District and Pennines and is exposed as a series of isolated outcrops, such as at Shap Fells, and encountered in boreholes below both the

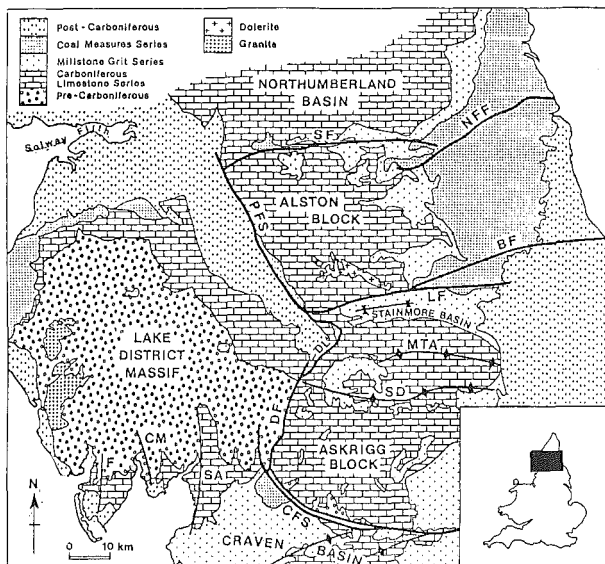


Figure 2. Geological Map of the Askrigg Block (SF, Stublick Fault; NFF Ninety Fathom Fault; PFS, Pennine Fault System; BF, Butterknoll Fault; LF, Lunedale Fault; DL, Dent Line; MTA Middleton-Tyas Anticline; SD, Stockdale Disturbance; DF, Dent Fault; CFS, Craven Fault System; F, Furness Fault; CM, Cartmel Monocline; SA, Silverdale Anticline). Reproduced by permission of the Geological Society from 'The Dent Fault System, northern England - reinterpreted as a major oblique-slip fault system' by Underhill, J.R., *et al.*, in the Journal of the Geological Society, volume 145, 1988).

East of the Dent Fault, the broad plateau summit areas and the continuous exposure of certain facies along the sides of the individual dales indicate that the rocks are nearly horizontal with only a slight regional dip to the east (Fig. 3). The lithological succession which forms this part of the Pennines is dominated by sedimentary rocks of Lower Carboniferous (Dinantian) age with outliers of Upper Carboniferous (Namurian) rocks forming the higher parts of the fells (Ramsbottom, 1974a, 1974b). The basal beds are a thick sequence of fluvial conglomerates, to which the formational name Sedbergh Conglomerates has been proposed (Burgess, 1986) and which lie unconformably on the Silurian strata in the Rawthey valley on the western side of the Dent Fault System (Fig. 3). They thin to both the north and south of the present outcrop in the Rawthey valley and it has been suggested that they are laterally restricted alluvial fan sediments deposited in a pull apart at a transtensional releasing bend in the Dent Fault at an early stage in its development (Underhill, *et al.*, 1988).

Within the Carboniferous rocks, there is a complex range of cyclothem lithofacies from deep water bioclastic limestones through fluvial and deltaic sandstones to coal seams (Fig. 3) (Ramsbottom, 1974a). This lithological repetition is well illustrated in upper Wensleydale and it is this sequence which Phillips (1836) proposed as his classic reference section to show the full stratigraphic complexity and to which he gave the name Yoredale Series.

The various cyclothems identified within the Dinantian are usually named after the limestones which form marker horizons within the sequence (Fig. 3), since they are laterally extensive and easy to correlate by their distinctive fossil assemblages (Hudson, 1924 ; Ramsbottom, 1974a). However, it has also been shown that a limestone in one area may have to be correlated with three limestone bands in an adjoining dale, thus illustrating the complex lithostratigraphic relationships which exist both across and up the sequence (Ramsbottom, 1974a).

The sandstone units usually begin with the formation of thinly bedded rippled sandstones which eventually grade upwards into massive cross-bedded sandstones (Moore, 1958). The upper part of the sandstones may be a quartz arenite (ganister) which formed as a palaeosol (Percival, 1986) following emergence of the delta top and the establishment of a land surface and associated vegetation cover which is preserved as coal seams within the sequence. Coarse grained cross-bedded

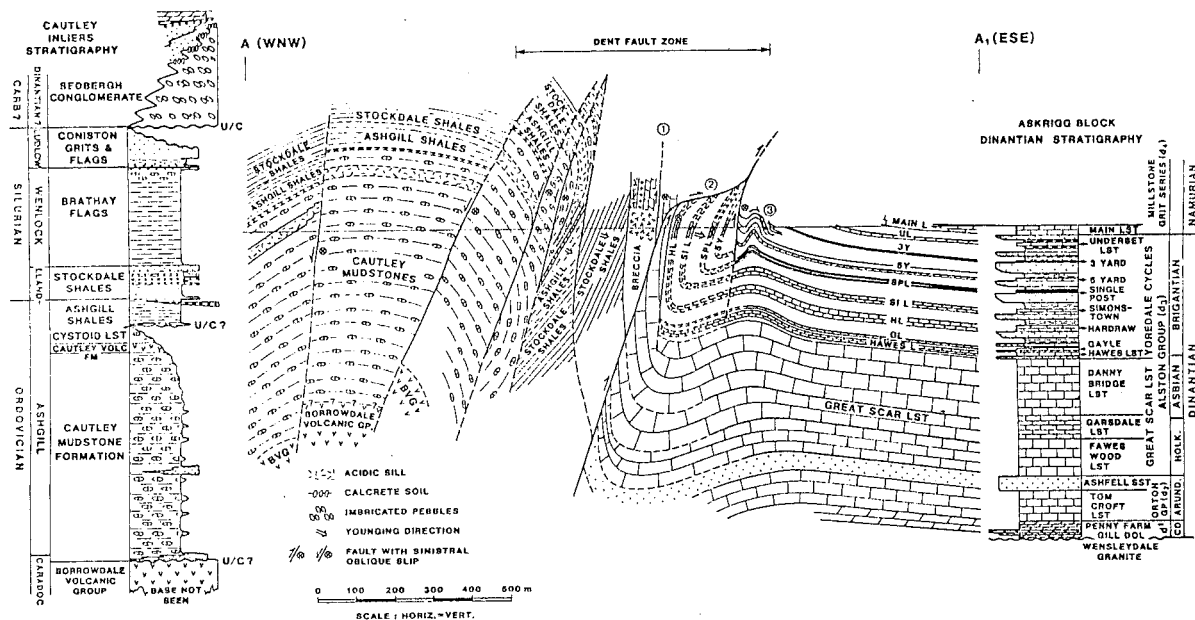


Figure 3. Summary stratigraphy on either side of the Dent Fault System, showing the chronostratigraphy for the Askrigg Block and the eastern Howgill Fells (from Underhill, *et al.*, 1988). Reproduced by permission of the Geological Society from 'The Dent Fault System, northern England - reinterpreted as a major oblique-slip fault system' by Underhill, J.R., *et al.*, in the Journal of the Geological Society, volume 145, 1988).

lenticular sandstones will reflect either palaeochannels cut into delta sediments as the distributaries changed their courses (Rowell and Scanlon, 1957a; Moore, 1958) or sand waves formed in a shallow marine environment (Brenner and Martinsen, 1990).

As can be seen in the published Geological Survey maps of the area, the sedimentary rocks are coded by age rather than lithofacies. For example, all Dinantian rocks are marked in blue which leads to perception that the sequence is dominated by limestones which is misleading, as these make up only a small fraction of total outcrop. This can be clearly seen on the lithological map which has been compiled for a part (SD 79) of the area (Fig. 4). Chronological divisions have been ignored and the major sedimentary facies represented instead. The limestones occur as narrow bands along hillsides which are predominantly composed of mudstones with inter-bedded sandstones (Rowell and Scanlon, 1957a).

The upper part of the fells (Fig. 4) are composed of lithologically similar facies, but the limestones are replaced by thick sequences of mudstones with occasional sandstone which become much coarser (grits) towards the fell summits. These are younger than the Yoredale Series of Dinantian age, and are defined as the lower units of the Millstone Grit Series of Namurian age (Rowell and Scanlon, 1957a, 1957b; Ramsbottom, 1974b) and reflect an increase in the amount of fluvio-deltaic clastic sediment input across shelf seas (Leeder and Strudwick, 1987; Leeder and McMahon, 1988; Gawthorpe, *et al.*, 1989).

The major geological structures are defined by the Σ -shaped fault patterns first identified by Sedgwick (1835a) and which divide this part of the Pennines into a series of east-west orientated 'blocks' and 'basins' (Fig. 2) associated with crustal stretching during the Carboniferous (Ramsbottom, 1974a, 1974b; Besly and Kelling, 1988; ; Underhill, *et al.*, 1988; Gawthorpe, *et al.*, 1989). The two major blocks, the Alston Block (Trotter and Hollingworth, 1928) and the Askrigg Block (Hudson, 1938) (Fig. 2), are present day upland areas and include many of the major summits of the Pennines.

History of Quaternary Research

Numerous lines of geological evidence, particularly till stratigraphy, striations, surface erratics and streamlined landforms have been employed in the interpretation of Quaternary events in Northern England. There have been many

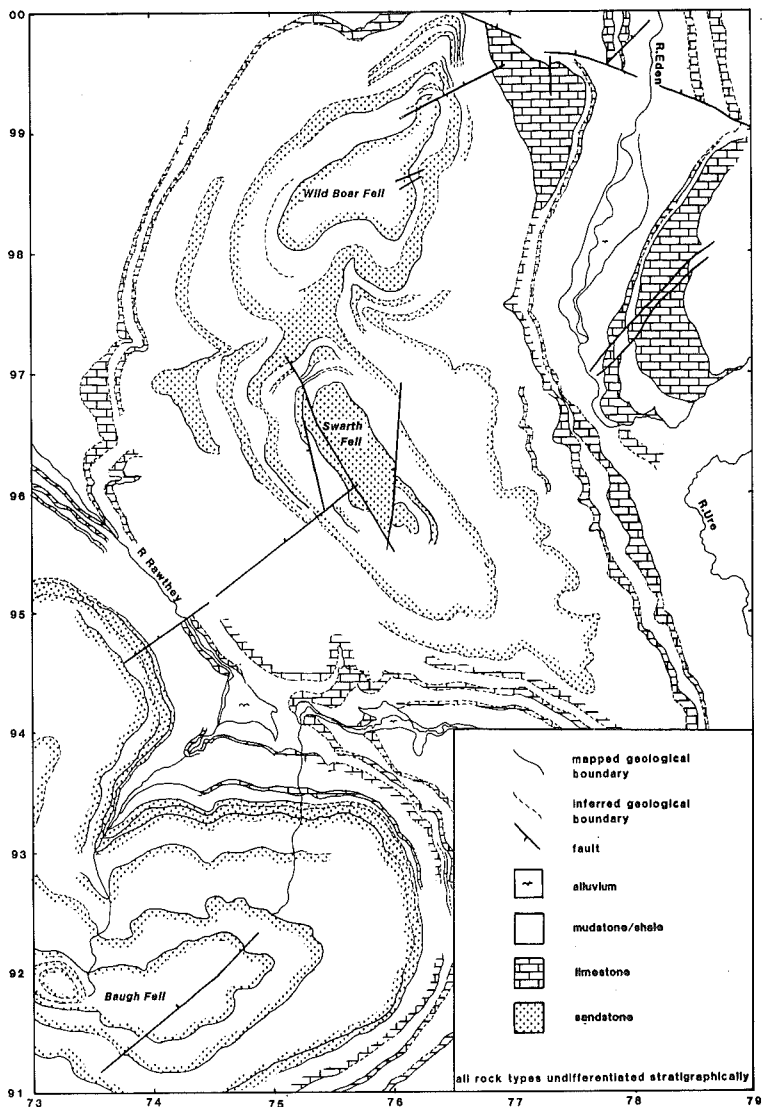


Figure 4. Lithological map of the area of OS area SD 79 to show extent of mudstones/sandstones in comparison to the limestone units (derived from information on Geological Survey maps; cf Dakyns, *et al.*, 1891; Rowell and Scanlon, 1957a, Figure 1).

key contributions to the Quaternary research of this part of the Pennines reflecting a continuing interest in the area since the last century (Table 1).

Early papers, including the Memoirs of the Geological Survey, concentrated on the interpretation of the glacial sequence, particularly with respect to ice sheet source areas, location of ice divides and thickness of the ice cover (Goodchild, 1875; Aveline and Hughes, 1888; Dakyns, *et al.*, 1890, Dakyns, *et al.*, 1891). All of these features were interpreted from a study of the pattern of former ice flow direction derived from striations and erratics (Fig. 5), although the evidence often conflicted, with ice flow directions apparently both down and up valleys

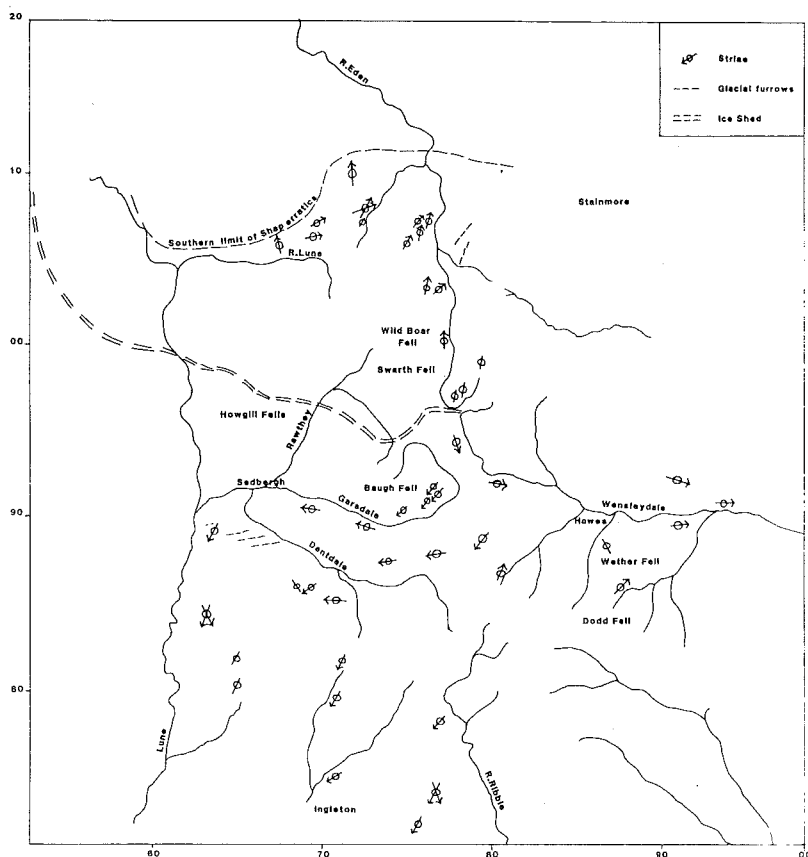


Figure 5. Ice flow directions in the Western Pennines (derived from Goodchild, 1875 and Dakyns, *et al.*, 1891).

<u>Early Workers:</u>	1872	Dakyns first discusses the evidence of land ice
		Tiddeman - regional account of N Lancashire/S. Askrigg Block
	1875	Goodchild - first detailed account of western Pennines/Vale of Eden
	1887	Goodchild - first ideas on glaciological processes
	1890/1	publication of the regional memoirs of the Geological Survey
	1894	Carvill Lewis - first attempt to map extent of last British ice sheet
	1909	Marr and Fearnside - regional account of the Howgill Fells
<u>Post Kendall:</u>	1902	Kendall - interpretation of meltwater channels as lake overflows
	1924	Kendall and Wroot - regional account of Quaternary of Yorkshire
	1926	Raistrick - glaciation of Wensleydale
	1927	Raistrick - interpretation of glacial retreat stages in Wensleydale
	1929	Trotter - examination of glacial geology of Vale of Eden/Alston Block
	1931	Hollingworth - glacial geology of western Vale of Eden
	1932	Trotter and Hollingworth - summary of work in Vale of Eden
	1933	Raistrick - summary of glacial activity in west Yorkshire
	1934	Raistrick - correlation of retreat stages across Pennines
		Hollingworth - notes periglacial features
	1935	King - examined drainage diversions in Wensleydale
<u>Recent Research:</u>	1952	Rowell and Turner - corrie glaciation in Mallerstang
	1955	Walker - radiocarbon date from Lunds
	1959	Manley - Lateglacial local glaciation in NW England
	1966	Gunson - pollen sites in western Pennines
	1967	Moulson - regional geomorphology of the Upper Lune
	1968	Beaumont - detailed investigation of tills in County Durham
	1969	Vincent - regional study of Alston Block
	1971	Huddart - investigation of Quaternary of Carlisle plain
	1974	Waltham - review of karst in Askrigg block
	1976	King - regional account of geomorphology of northern England
	1978	Letzer - detailed study of the southern Vale of Eden
		Pounder - detailed study of periglacial fluvial landforms in Swaledale
		Gascoyne - first U/Th dates from speleothems
	1980	Rose - landforms in upper Swaledale
	1981b	Boardman - detailed study of NE Lake District
	1981	Harvey <i>et al.</i> - postglacial fluvial activity in the Howgill Fells
	1985	Johnson - review of glaciation in western Pennines
	1989	Eyles and McCabe - study of Irish Sea Basin

Table 1. Important contributions to the Quaternary geology of northwest England.

(Goodchild, 1875). This was explained as being due to the ability of the ice to move in different directions at varying depths within the ice sheet, with the lower ice being constrained by topography and flowing down the valleys, but with the upper ice flowing radial from a local ice centre (Goodchild, 1875, 1887). Only erratics of local provenance were identified in the area of the upper Wensleydale and the adjoining regions indicating that the ice was local in origin with an ice divide/centre across Wild Boar Fell/Baugh Fell (Goodchild, 1875, 1887; Dakyn, et al., 1890, 1891).

Carvill Lewis (1894) noted that glaciers in northern England had descended the numerous valleys extending beyond the mountains to the surrounding lowland areas (Fig. 6) with the "most complex glacial movements in England occurring in the mountain region about Nine Standards where local glaciers met and were overpowered by the greater ice sheet coming down from Cumberland" (Carvill Lewis, 1894; p.13). As with the other workers at this time, Carvill Lewis thought that from the position of lateral moraines, the ice did not cover the high ground between Wensleydale and Stainmore and with only a valley glacier extending to Reeth in Swaledale leaving the rest of the valley ice free (Fig. 6). A final paper which may be included within the initial papers on glaciation is the paper by Marr and Fearnside (1909) for the Howgill Fells, whose conclusions showed general agreement with the earlier work, pointing to the existence of an independent but ineffective ice centre over these fells with ice flowing outwards into the surrounding lowlands.

In their review of the geology of Yorkshire, Kendall and Wroot (1924) noted that at the maximum of the last ice sheet, the upland area at the head of Wensleydale was overrun by ice from the Howgill Fells and Ravenstonedale and that the area to the south of Wensleydale, around Blea Moor and Pen-y-ghent also provided a southern source area for tributary ice to the main Wensleydale ice and to another ice mass in Wharfedale and Littondale. Developing the model of ice sheet retreat used to explain the meltwater channels in north-east Yorkshire (Kendall, 1902), they suggested that there must have been many glacial lakes formed during ice retreat, although no specific evidence or details are given for Upper Wensleydale.

The geological evidence within the Vale of Eden and the surrounding uplands, with the complicated interpretations which were required to explain ice movement, was re-examined in the eastern part of the Vale of Eden and the Alston

Figure 6. Ice limits and flow lines of the last ice sheet in northern England (from Carvill Lewis, 1894).

Block by Trotter (1929) and the western part of the Vale of Eden and the surrounding uplands of the Western Pennines were reported in Hollingworth (1931), who produced the first systematic account of the drumlins in the Vale of Eden. Both authors built upon the interpretations made by Goodchild (1875, 1887) but they were more confident at developing a stratigraphy of glacial events based on detailed investigations and correlation of Pleistocene sediments from a number of lowland areas in northern England (Trotter and Hollingworth, 1932).

At the same time as the lowland areas to the north of the Askrigg Block were being mapped, attention was also being focussed on the sequence of glacial events within the Yorkshire Dales, particularly with respect to the nature of ice retreat (Raistrick, 1926). The lack of any far travelled pebbles again confirmed the presence of only local ice with a direction "radial from the Mallerstang valley or the Wild Boar Fell massif" (Raistrick, 1926; p.369) with an ice divide in Mallerstang. The sequence of terminal moraines and 'overflow channels' identified in the Yorkshire Dales allowed a model of glacier retreat to be proposed which involved the gradual retreat of the ice margin of each of the valley glaciers westwards with lakes impounded behind the terminal moraines (Raistrick, 1926, 1927, 1933).

More recent studies have continued the glacial theme with regional accounts of the Lune valley (Moulson, 1957), Alston Block (Vincent, 1968), the southern Vale of Eden (Letzer, 1978) and part of upper Swaledale (Rose, 1980) (Table 1). Letzer (1978) marked a change in style by completing detailed field mapping at 1:10,000 making accurate data on drumlin form available for the first time. Previous studies on drumlins had usually extracted data from topographic maps (cf Doornkamp and King, 1971; King, 1976) but such an approach was critically reviewed by Rose and Letzer (1975) who found that in most cases, such map-based studies produced drumlin distributions which showed little similarity to the drumlin pattern determined by detailed field mapping. It was also demonstrated that the glacially streamlined landforms could be described as part of a continuum of megadrumlins, drumlins and superimposed drumlins (Rose and Letzer, 1977; Letzer, 1978, 1981, 1987; Rose, 1987, 1989b).

More recently, attempts have been made to interpret the geological evidence for the pattern of deglaciation in the Irish Sea Basin employing the concept of rapid marine ice sheet disintegration (marine drawdown) to explain the rapid collapse of the late Pleistocene ice sheets (cf Denton and Hughes, 1981; Hughes, 1987). This

has implications for the interpretation of ice movement in the surrounding uplands. Eyles and McCabe (1989a, 1989b) have interpreted the sedimentological and geomorphological evidence, from the terrestrial areas surrounding the Irish Sea and from the sea floor itself, to produce a model for the retreat of a marine ice margin within the basin and explain the development of the different drumlin fields around the Irish Sea Basin as the result of ice being rapidly drawn out from the source areas to compensate for high calving rates at the ice margin.

A second more restricted glacial event (Loch Lomond Stadial), which formed local glaciers in many upland areas, is widely acknowledged to have followed the deglaciation of the last (Dimlington Stadial) ice sheet (*cf* Gray and Coxon, 1991). Geomorphic evidence, in the form of terminal moraine ridges, for the occurrence of these local ice masses in northern England was discussed by Manley (1959) but this was not based on detailed mapping of landforms. Some of the landforms had been noted by early workers (Goodchild, 1875; Dakyn, *et al.*, 1891) but generally interpreted as associated with ice sheet downwastage and not a separate glacial event. In Mallerstang, Rowell and Turner (1952) mapped a series of ridges and mounds as evidence for the former existence of local glaciers. Their genesis has since been questioned (King, 1976; Letzer, 1978), but detailed mapping of these has shown that most of them can be better explained as features associated with large scale slope failures (Mitchell, 1991, this guide). This period of local glaciation is confirmed to be of Loch Lomond Stadial age by pollen analysis from within the moraines at Combe Scar (Gunson, 1966, this guide).

Research over the last forty years has shown that the present landscape of the northern Pennines is the result of the operation of different geomorphological processes and not only due to the impact of Pleistocene ice masses (Table 1). It is also the period marked by the development of a more rigorous chronology into which geomorphic events may be placed (Rose, 1989a; Bowen, 1991). Studies into periglacial phenomena (Tufnell, 1969, 1985), periglacial fluvial processes (Pounder, 1981, 1985) and the operation of fluvial activity (Harvey, 1985; Harvey, *et al.*, 1981, 1984) have all shown the complexity of landform development in the area.

The nature of environmental change, with particular reference to palynology (Fig. 7), has been the subject of research papers on the Lake District (Pennington, 1974, 1977, 1978), the Alston Block (Turner, 1984) and in the western Pennines (Gunson, 1966; this guide). The presence of extensive limestone outcrops

in the southern part of the Askrigg Block has also meant that this area has been an important area for karst geomorphology (Sweeting, 1972, 1974; Waltham, 1974; Gunn, 1985) and the use of speleothems for dating Pleistocene climatic change (Ivanovich and Harmon, 1982; Gascoyne, *et al.*, 1983; Schwarcz, 1989; Schwarcz and Gascoyne, 1984).

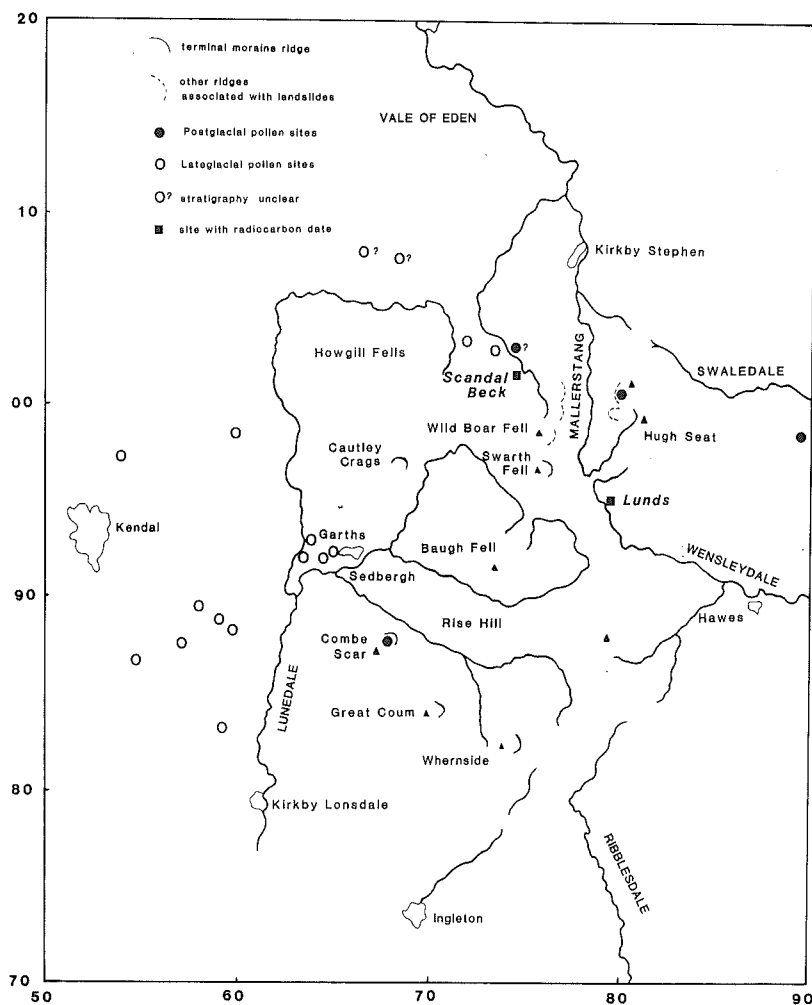


Figure 7. Pollen sites in the Western Pennines (data derived from Walker, 1955a, 1955b; Gunson, 1966, unpublished; Letzer, 1978; Rose, 1980; Carter, *et al.*, 1982).

Chronology of Glacial Events

More recent studies of the British Quaternary have benefited from the development of a more rigorous chronostratigraphy for the British Isles into which the glacial events may be placed (cf Bowen, et al., 1986; Ehlers, et al., 1991). Chronological control of glacial events in the Western Pennines is however very limited and there are no sites where dates are available for the glacial events for which there is geomorphic evidence. The ice sheet glaciation is assumed to have occurred during the Dimlington Stadial (c. 26-13 ka) (Rose, 1985, 1989a) on the lateral correlation of till units across northern England to the proposed type site at Dimlington where dates of c. 18,500 yrs BP have been obtained from moss in lake sediments (Dimlington Silts) below the tills (Penny, et al., 1969; Rose, 1985, 1989a; Catt, 1991a, 1991b).

Limited evidence for glacial events older than the late Devensian (Dimlington Chronozone) Substage is recorded in northern England (cf Bowen, et al., 1986; Rose, 1989a)(Table 2). At the proposed type site, at Dimlington, and at other sites along the eastern coast, there are outcrops of a lithofacies known as the 'Basement Till' (Catt and Penny, 1966) which underlies the Dimlington Silts and which is interpreted as of Wolstonian age (Bowen, et al., 1986; Catt, 1987; Catt and Digby, 1988; Rose, 1989a). In the eastern Lake District, there is a deeply weathered till and associated palaeosol which underlies Dimlington Stadial till, but the age is far from certain and can only be regarded as pre-Devensian (Boardman, 1985, 1991). At Scandal Beck (NY 743025), near Ravenstonedale (Fig. 7), there is a site exposed in the bank of a northerly flowing tributary of the River Eden which shows a complex stratigraphy of glacial sediments which overlie and incorporate organic material, interpreted as of last interglacial (Ipswichian) age on the pollen evidence (Carter, et al., 1978; Letzer, 1978, 1981).

Speleothems from the limestone areas of the southern part of the Askrigg Block give an important database for long term environmental change, providing information regarding the number and timing of possible episodes of glaciation going back to the Middle Pleistocene (Table 2). Whilst the timing of earlier glacial events remains uncertain due to a lack of other types of evidence, the U-Th dates obtained from the speleothems clearly show that the last major ice sheet glaciation occurred during the Dimlington Stadial (Gascoyne, et al., 1983; Schwarcz, 1989; Schwarcz, et al., 1982).

		oxygen isotope stage	Eastern Lake District ¹	Vale of Eden ²	Study Region ³	Yorkshire Dales ⁴	Alston Block ⁵	Vale of York ⁶	Holderness ⁷
HOLOCENE		1	peat / fluvial sediments/ landslide debris; Ladray Wood palaeosol	peat / fluvial sediments; river terraces / alluvial fans	peat / fluvial sediments; river terraces / alluvial fans; landslide debris speleothems	peat speleothems	peat	fluvial sediments peat	fluvial sediments peat lake sediments
DEVENSIAN	Loch Lomond Stadial	2	Skiddaw scree and lateglacial greze lites/ proglacial ramparts/rock glaciers; Wolf Crag till & gravel/ cirque moraine ridges	alluvial fans gelifluctate (head) landslide debris	scree / gelifluctate; proglacial ramparts/ rock glaciers; landslides till and cirque moraine ridges	?	gelifluctate landslide debris	cover sand lake sediments	cover sand lake sediments
	Windermere Interstadial		organic muds	organic muds	landslides ? Lunds organic deposit	organic muds	?	organic muds	organic muds
	Dimlington Stadial		valley glacier moraine ridges/kames/ kettle holes; meltwater channels/ drumlins. Threlkeld till. Lobbs sand & gravels Mosedale gravel	meltwater channels; sand & gravels; erratic brains/ till (Scandal Beck) drumlins (Howgill Suite) drumlins (Bough Fell Suite)	landslides, valley glacier moraine ridges meltwater channels/kame drumlins (Wild Boar Fell Suite) drumlins (Bough Fell St.)	valley glacier recessional moraines; meltwater channels/sand & gravel 'Main Dales' till (newer drift)	till	lake sediments (Lake Humber) York-Esrick moraines drumlins	Withernsea Till sand and gravels Skipssea Till Dimlington silts
	Middle	3			speleothems	speleothems		Oxbow silts (mammoth remains)	
	Early	4 5a-d							Sewerby blown sand
IPSWICHIAN			Troutbeck palaeosol	Scandal Beck organic deposit	speleothems	Victoria cave fauna (hippopotamus remains) speleothems		'older gravel' (?)	Sewerby beach gravels
(Earlier geological events)		6 of seq.	Thornsgill weathered till	Scandal Beck lower till	speleothems in interglacials	'Maximum Dales' till		'Older Drift' (?)	Basement Till

¹ Beardman 1981, 1985
Sissons 1980a

² Letzer 1976, 1981

³ Walker 1955
Gascoyne et al 1983
this study

⁴ Ralstrick 1926
King 1975
Penny 1974
Rose 1985
Gascoyne et al 1983

⁵ Vincent 1968
Tufnell 1985

⁶ Gaunt 1981

⁷ Catt 1987
Catt and Digby 1986
Rose 1985

Table 2. Correlation chart of Pleistocene landforms and sediments with time in northern England (from Mitchell, 1991).

The earliest dates associated with deglaciation are from the southern part of the Askrigg Block, where speleothems have given dates of 14.6 ± 0.6 and 14.7 ± 0.7 ka (Gascoyne, *et al.*, 1983) and which suggest that by this time, the area was not only deglaciated, but that vegetation had started to grow on the surface. Comparable or earlier dates for organic sedimentation in the Lake District and the North York Moors give ^{14}C dates of a similar age for deglaciation, although due attention must be given to standard errors and hard water contamination. Furthermore, if the proposed calibration curves for U/Th and ^{14}C are correct (Bard, *et al.*, 1990), then this U-Th date cannot be directly compared to ^{14}C dates. In this case the ^{14}C dates should be amended to make a 14.5 ka determination of 18.2 ka which clearly illustrates current difficulties of giving a meaningful chronostratigraphy for late Devensian events.

Following the disappearance of the ice sheet from the British Isles, there is widespread evidence from the mountain areas in northern England (Table 2) and elsewhere, for a more limited redevelopment of the glaciers during the Loch Lomond Stadial (11-10 ka BP) (Rose, 1989a; *cf* Gray and Coxon, 1991).

GEOMORPHOLOGICAL MAPPING : AN INTRODUCTION

W.A. Mitchell

Production of comprehensive geomorphological maps has been pioneered by a number of European geomorphologists (cf Demek & Embleton, 1978; Barsch & Liedtke, 1980; Embleton and Verstappen, 1988). Such mapping has often been carried out as part of a national survey and usually includes data on geological structure, topography, landform genesis, past processes and landforms, and hazards. However, geomorphological maps may also be constructed for a specific purpose and only record information directly relevant to a particular research project.

Geomorphological mapping, specifically orientated to the landforms and sediments associated with Pleistocene glacial and periglacial activity, has only become an important research tool during the last thirty years. From this more exact database a new understanding and original interpretation of many issues of Quaternary geomorphology has been achieved. Examples of this work include papers on the origin of meltwater channels and their significance in reconstructing ice wastage patterns (Sissons, 1958, 1960, 1961; Young, 1974; Gray 1991), the sequence of raised marine shorelines and their significance to processes of isostatic rebound and deglaciation (Rose, unpublished, Sissons, *et al.*, 1966) and the extent of the Loch Lomond Stadial glaciers in Scotland and northern England (Sissons, 1974b, 1977a, 1977b, 1979b, 1980a; Gray, 1982; Ballantyne, 1989; Gray and Coxon, 1991). This is also the case in the construction of geomorphological maps of slope stability required by engineering geologists (Conway, 1974; UNESCO, 1976) and for planning purposes (de Mulder, 1989).

This approach has also been used in research in northern England in areas adjacent to the study area, for example in the southern Vale of Eden (Letzer, 1978), in part of Swaledale (Rose, 1980) and in the north-east Lake District (Boardman, 1981). A further area where detailed field mapping has proved to be extremely profitable has been in the study of mass movement features along the southern coast of England (Brunsdon and Jones, 1972). However, in all cases the technique is most effective in regions, where the landforms have a discrete recognisable shape and have not been modified by subsequent processes, such as

the drumlins which are found in many parts of the Western Pennines.

Detailed geomorphological mapping is achieved by the accurate recording of the different landforms by walking across the area and plotting the data on 1:10,000 scale maps. Geomorphological mapping requires the precise location of the landform boundaries and this is done with reference to features located on the base map. Features are mapped by standing on them and traversing them, as distant mapping can introduce inaccuracies due to perspective which may lead to distortion of the actual shape of the landform being delimited on the map and would result in incorrect measurements being recorded for morphometric analysis. The landforms are identified on the map by the use of a set of symbols which have been developed to suite the landforms found within the study area (Fig. 8). The basis of the classification of landforms is found in the early work of J. B. Sissons (Sissons, 1958) and subsequently developed in the mapping programmes carried out by Rose (Rose, unpublished, 1980, 1981) and Letzer (Letzer, 1978). For example, a geomorphological map will allow detailed definition of the complexity of drumlin forms in an area (Fig. 9) which can then be used for detailed morphometric analysis of the form to provide information on the former subglacial environment of the last Dimlington Stadial ice sheet under which they were formed (Mitchell, 1991).

The definition of specific constructional and erosional forms is based primarily on recognisable breaks of slope, which give a distinctive outline and areal expression to specific features. In terms of geometry, there are a small number of identifiable forms. Ridges, mounds and hollows together with distinct linear breaks of slope, which define different slope gradients, can be regarded as the main morphological output from the operation of many different process response systems. However, because they are common forms to many geomorphic systems, they cannot be used to identify the operation of specific processes.

Only when the landforms are placed into their environmental setting and are found in juxtaposition with distinctive geometric forms which are characteristic of a specific geomorphic system, can such general forms be specifically defined and used in the establishment of a genetic interpretation of the landforms. For example, constructional ridges are formed by the operation of glacial, glaciofluvial, periglacial and marine processes. Similarly, protalus ramparts, terminal moraines and rock glacier ridges can be distinguished with confidence only when the operator is able to assess the environmental situation. In such situations the

GLACIAL LANDFORMS AND SEDIMENTS



Drumlin with crestline
and highest point



Till



Drift mound with highest point



Moraine ridge
with crestline

GLACIOFLUVIAL LANDFORMS



Meltwater channel

PERIGLACIAL LANDFORMS AND SEDIMENTS



Sorted/non-sorted circles



Gelifluction lobes



Sorted/non-sorted polygons



Stone banked lobe

FLUVIAL LANDFORMS AND SEDIMENTS



River channel



Alluvium



Alluvial fan

MASS MOVEMENTS



Rotational (in early mapping)



Debris flow



Rotational (in later mapping)



Levées /compression ridges



Tension gash



Areas of boulders



Marsh



Mound



Bedrock

Figure 8. Symbols used in geomorphological mapping; (on the original field slips, lithological data are coloured as indicated) (based on J. Rose, unpublished, Letzer, 1978 and Mitchell, 1991).

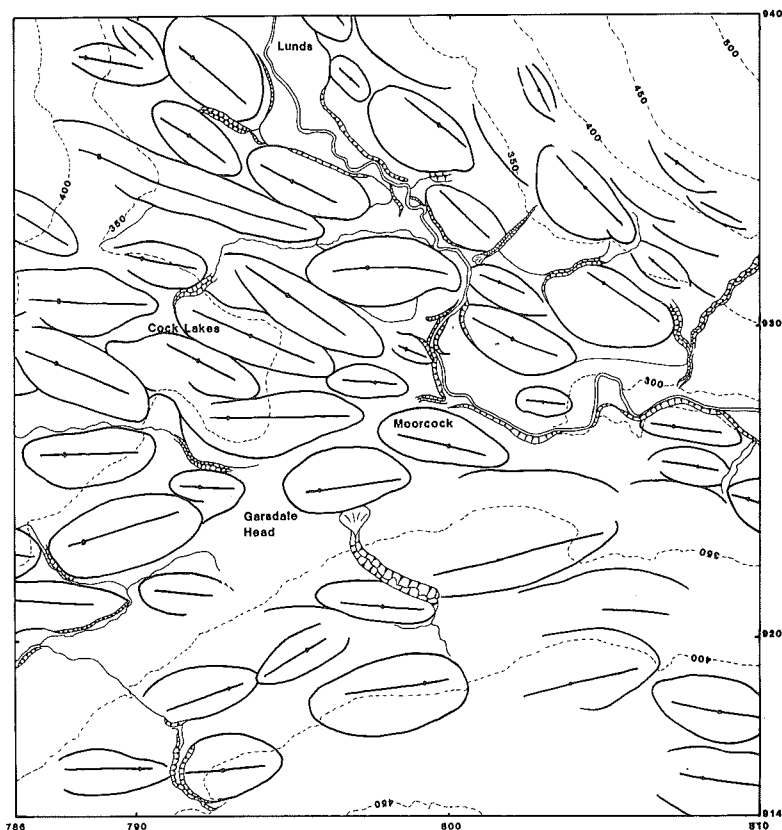


Figure 9. Example of a geomorphological map showing drumlins around Moorcock Inn at Garsdale Head (from Mitchell, 1991). Scale and orientation are given by National Grid coordinates.

explanation of origin will reflect the relationship of the ridge to the talus/debris supply and its position relative to the hillslope. However, it must be appreciated that such ridges may be polygenetic in origin and have been formed through the operation of a talus derived landform continuum (Shakesby, *et al.*, 1987).

The aim is therefore not just to create an accurate record of the morphological expression in an area, but to place it within an appropriate genetic process-response definition. A geomorphological map should therefore provide an important means of expressing the data on the landforms of an area in an accurate

spatial manner (Fig. 9). This is very important because it allows the relationship between different landforms to be assessed and measured. Furthermore, only by completing a detailed survey of an area can the full detail of the landform assemblage be appreciated. When the complete succession of landforms is known a chronology of geomorphic development may then be inferred.

Although the emphasis is on form, information on the underlying geology is of importance. The location of rock outcrops shows where there is no deposition of superficial deposits. It also provides the materials on which geomorphological processes must operate and controls specific processes through the operation of material stress/strength relationships, such as bedrock deformation or deep seated rotational slope failures. Also, the overall expression of the landscape is often shown to have a strong lithological and structural control as can be seen within the Western Pennines, in the different landscapes of the Howgill Fells composed of Silurian rocks and Pennines of Carboniferous strata.

The geomorphological map must also include information on the sediments of which the landforms are composed. The sediments are important for two reasons. Firstly, they provide information which might be important in determining the origin of the feature in terms of depositional environment. This will allow landforms of similar morphology but composed of different sediments, to be classified in terms of genesis. Secondly, if there are suitable materials for dating within the sedimentary sequence, an attempt can be made to establish a chronology of geomorphic events.

DIMLINGTON STADIAL ICE SHEET IN THE WESTERN PENNINES

W.A.Mitchell

The last ice sheet to cover the north of England was part of a much larger ice mass which inundated most of Scotland, Ireland and extended into the English Midlands and the southern part of Wales during the Dimlington Stadial (Rose, 1985; Bowen, et al., 1986; Catt, 1991a, 1991b). Geological evidence from striations, erratics and drumlins have shown that this ice sheet was the combined accumulation of ice from a number of different source areas (Johnson, 1985; McCabe, 1985; Sutherland, 1991). Upland areas in northern England, such as the Lake District and Pennines, were able to generate a local ice cover which was able to expand into the surrounding lowland areas at times when there was a reduction in influence from Scottish ice, particularly from the Southern Uplands. Yet beyond a general regional understanding of glacial events (Mitchell and Rose, this guide), there have been few detailed studies of the upland areas to improve the resolution of this general model. The Western Pennines is an area for which new information is now available (Mitchell, 1991) and which can be correlated to a sequence of events proposed for the southern part of the Vale of Eden (Letzer, 1978, 1981, 1987). This chapter is based on detailed mapping of the drumlins, superimposed drumlins and drift tails across 350 sq. km of the Western Pennines which has allowed former ice flow directions, the identification of ice streams and ice divides, aspects of the palaeoglaciological character of the ice sheet and the pattern of deglaciation to be established (Mitchell, 1991). This chapter will briefly introduce each of these topics.

Drumlin Distribution

A major line of evidence in the reconstruction of former ice directions is the mapping the streamlined glacial bedforms (drumlins, superimposed drumlins and drift tails) which are well distributed throughout the area (Fig. 10). 952 drumlins have been mapped not only in the valley areas, but also on many of the interfluve areas, up to 667m. This suggests that the flow pattern reflected by the trend of the drumlin long axes indicates that the former ice sheet across the area was sufficiently thick to be acting independently of topographic control. The distribution also shows that there is a central area across the high fells, from Wild

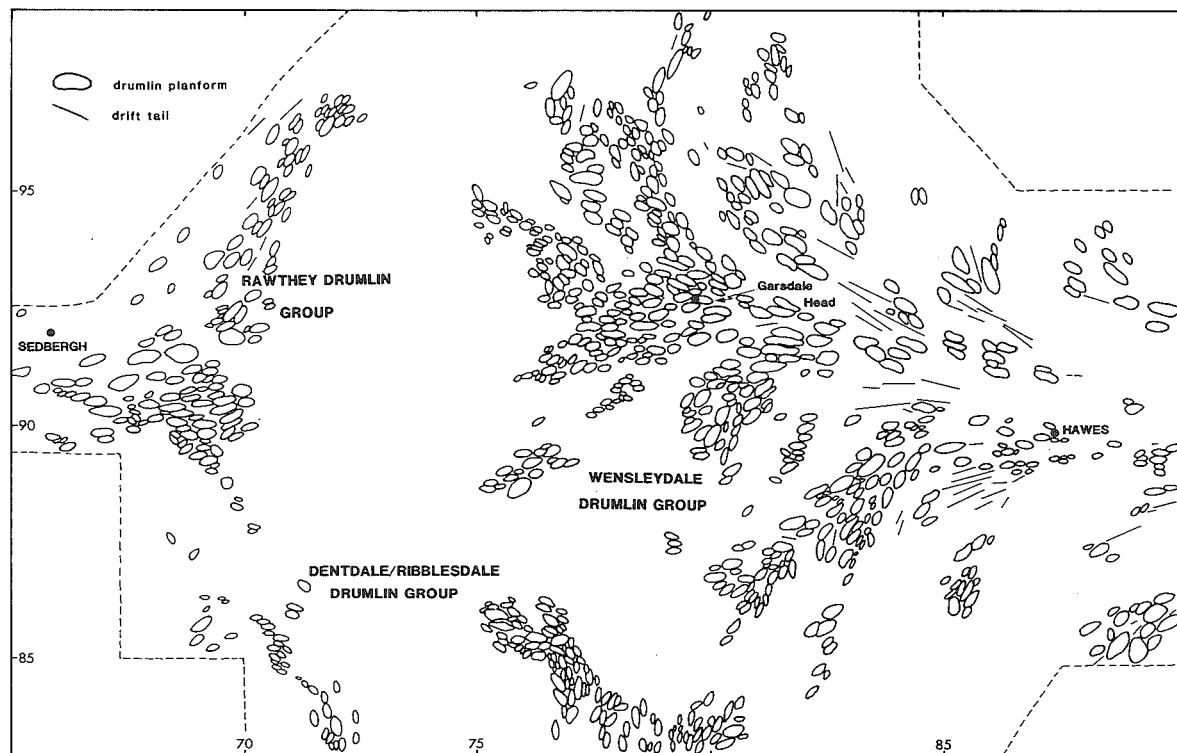


Figure 10. Pattern of drumlins and drift tails in upper Wensleydale and adjoining regions (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

Boar Fell (SD 7598) in the north to Whernside/Great Coum in the south where there are no drumlins. This allows the drumlins to be divided into three groups: Wensleydale Group, Rawthey Valley Group and Dentdale/Ribblesdale Group (Fig. 10).

Detailed field mapping has provided data for morphometric analysis of drumlin shape with the following morphological variables being measured for each drumlin - length, width, area, orientation, elongation ratio, k and stoss distance ratio. Spatial and statistical analysis has been computed for each variable and the overall distribution and trends of drumlin morphology across the area have been determined and has formed the basis for many of the statements regarding the palaeoglaciology of the ice sheet (Mitchell, 1991).

Ice Flow Direction

In upper Wensleydale and the tributary valleys, former ice flow converged into the main valley from the interfluvial areas which define the present catchment of the River Ure (Fig. 10). In most instances, former ice flow was down valley, although in the eastern part of Garsdale, the continuous spatial pattern of drumlins indicates that ice flow was up valley. High level drumlins are found in Widdale Fell indicating ice flow from Dentdale into the Wensleydale catchment, and on the interfluvial areas between the northern tributary valleys. A continuous distribution of drumlins from the upper part of Ribblesdale across to Dent Head shows that ice flow for the Dentdale/Ribblesdale drumlin group was southwards. A similar pattern is found in Deepdale with ice flow east, then south into Kingsdale and these drumlins are included in this second group. The third group of drumlins has been mapped in the Rawthey valley and across the interfluvial area of Frostraw Fells between Garsdale and Dentdale (Fig. 10). These drumlins indicate southwesterly ice flow off Wild Boar Fell (SD 7598) and Baugh Fell (SD 7392) which coalesced with ice flowing down the lower parts of Garsdale and Dentdale to become part of the ice occupying the Lune valley and flowing off the western Howgill Fells and high ground of Whinell Beacon, towards the Lake District.

Superimposed drumlins within the area indicate that two distinct stages of ice flow direction are recorded at a number of localities, with the evidence from these drumlins showing that during the course of the glaciation/deglaciation, ice flow directions changed reflecting the relative importance of different ice source areas. (Fig. 11). For example, the original drumlins on Frostraw Fells suggest an initial dominant ice flow down Dentdale with ice able to diverge into the lower part

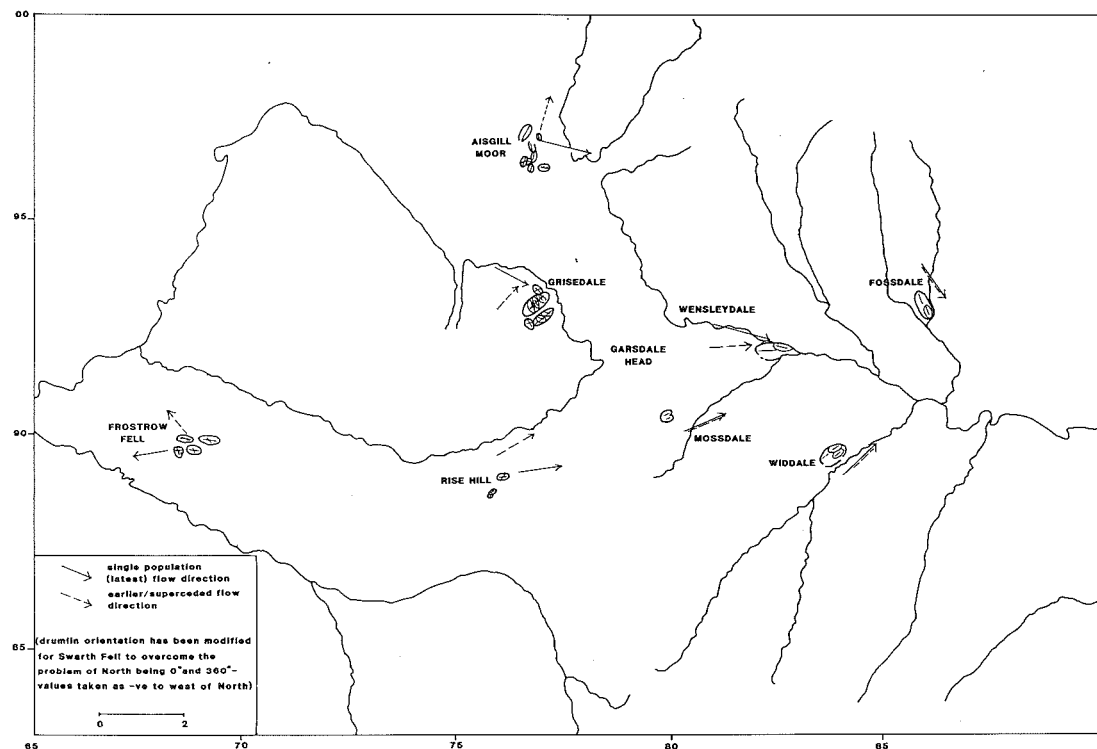


Figure 11. Areas within the Western Pennines where superimposed drumlin forms have been identified during field mapping (from Mitchell, 1991). Scale and orientation given by National Grid Coordinates.

of the Rawthey valley, with Rawthey/Garsdale ice being a secondary smaller flow. However, this pattern is found to change as the superimposed forms record a more dominant flow out of Garsdale. In Grisedale, large drumlins are orientated across the valley to the northeast, indicating ice flow off Baugh Fell (SD 7392) towards Mallerstang, which was superseded by ice flow down valley. A similar pattern is found on Aisgill Moor (Fig. 12). Superimposed drumlins on larger forms have also been identified in Wensleydale where they may reflect changing ice dynamics but with no change in flow direction. This information can be used to define two ice flow events. During the initial stage, ice flow was from an area between Baugh Fell and Great Coum (**Ice flow event 1**) (Fig. 12) and that this was succeeded by a second, later flow event (**Ice flow event 2**) (Fig. 13), by which time the source area had extended northwards to include Wild Boar Fell (Mitchell, 1991). This has implications to the morphology and dynamics of the ice divide in the Western Pennines.

Ice Streams

Ice streams are one of the most critical components of both marine and terrestrial ice sheets for meaningful palaeoglaciological reconstruction, since they reflect specific subglacial conditions, mass flux and drainage patterns within the ice sheet (Hughes, 1981, 1987; Hughes, *et al.*, 1985; Boulton, 1986; Bentley, 1987).

Convergent flow pattern and abrupt lateral termination of drumlins and/or erratic trains have been used in the identification and delimitation of sediment plumes/former ice streams of the Laurentide Ice Sheet (Andrews, 1987; Andrews, *et al.*, 1985; Shilts, 1982, 1985; Dyke and Morris, 1988; Dyke and Dredge, 1989; Dyke, *et al.*, 1989). In the study area, a convergent pattern of ice flow is clearly observed in the drumlin trends within the Wensleydale drumlin group and Vale of Eden towards Stainmore (Fig. 14). This is associated with the second ice flow event, and appears to conform to the pattern that would be expected from the accumulation area of an ice stream (Fig. 15). Although studies of present day ice streams show that many do not have an underlying trough (Bentley, 1987), Wensleydale clearly provides a major conduit to ease mass flux eastwards down the valley which would define a former ice stream approximately 70km in length and 10km in width (Mitchell, 1991). The convergence of flow indicated by the Stainmore/Lake District drumlin suite within Stainmore (Letzer, 1978) is also thought to reflect the position of a former ice stream (Fig. 15).

The interpretation of former ice streams from the geomorphic evidence is more

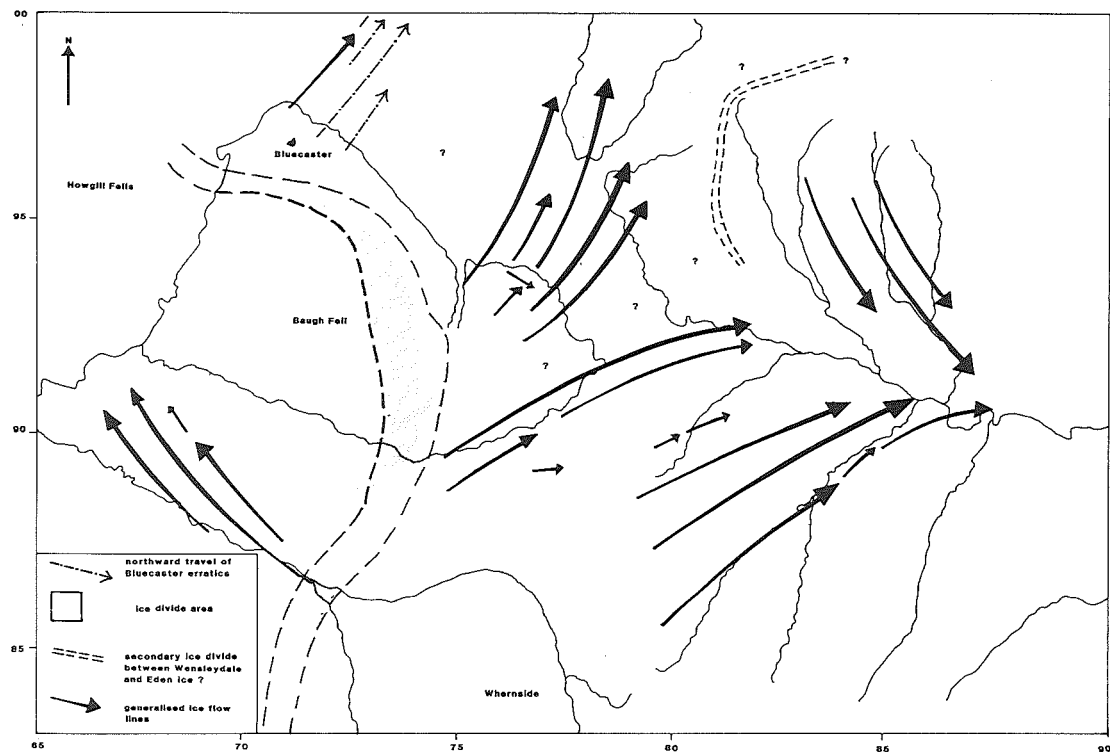


Figure 12. Iceflow directions and location of ice divides in Western Pennines during Ice Flow Event 1 (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

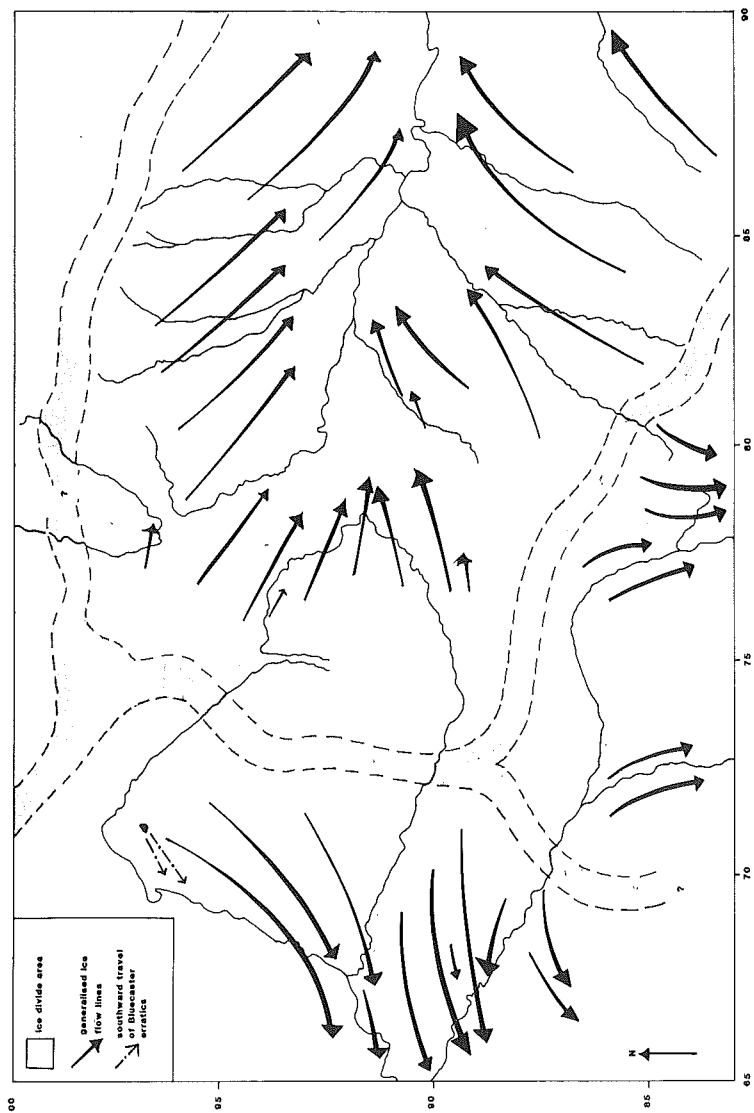


Figure 13. Ice flow directions and location of main ice divides in Western Pennines during Ice Flow Event 2 (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

equivocal for the other two drumlin groups, because they are less extensive within the presently mapped area (Mitchell, 1991). The underlying topography is more diverse to the south in Ribblesdale and west in Lunedale where the detailed drumlin distribution is unknown although complex flow is apparent, particularly where the ice moves south off the Askrigg Block (Johnson, 1985).

These former ice streams are a key morphological element in deciphering the dynamics and glaciological characteristics of the former ice sheets. This may be particularly the case, where rapid drawdown into an ice stream caused by increased or initiation of rapid ablation by calving at the ice margin, would cause the migration of the upper part of an ice stream into the centre of the ice sheet to maintain the flux balance (cf Denton and Hughes, 1981). If the ice stream migrated

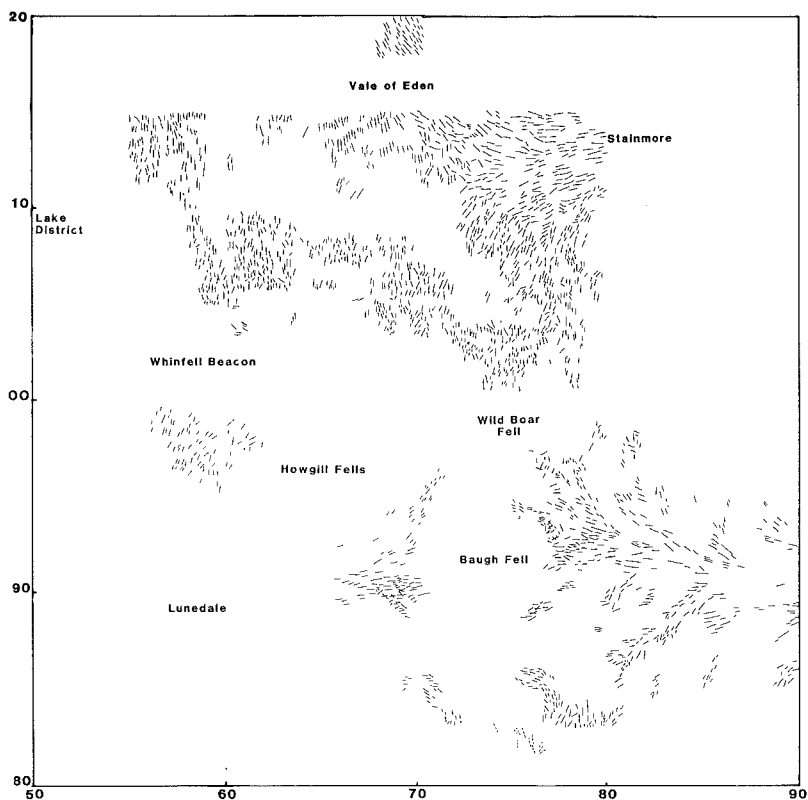


Figure 14. Ice flow pattern in southern Vale of Eden and Western Pennines as derived from drumlin long axes (derived from data in Letzer, 1978, Whiteman, 1981, M. J. Brunskill, unpublished, and Mitchell, 1991). Scale and orientation given by National Grid coordinates.

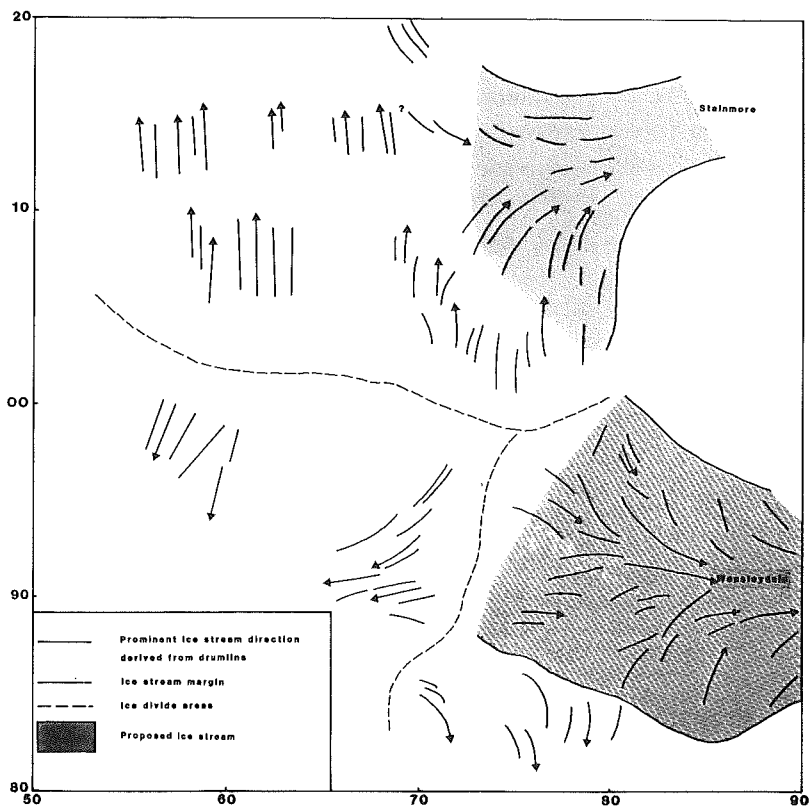


Figure 15. Proposed ice streams in upper Wensleydale and Stainmore (data derived from Letzer, 1978, M. J. Brunskill, unpublished and Mitchell, 1991).

all the way back to the ice divide, then it suggests that the ice sheet was in a state of critical collapse, unless there was exceptionally high accumulation to maintain the mass flux that the ice stream would require from a very small catchment area.

The model proposed by Eyles and McCabe (1989a) for glaciomarine sedimentation within the Irish Sea Basin involves the process of marine downdraw to generate ice sheet collapse. To maintain a marine ice margin, which would be unstable and subject to high ablation, there would have to be an increase in mass flux by increasing velocity to draw ice out of the accumulation zone. It is these increased velocities, to possible surge conditions, that are envisaged by Eyles and McCabe (1989a) to have formed the large drumlin fields in Ireland. Whilst this may explain the 'ice streams' directed towards the Irish Sea Basin, it does not explain the bedforms associated with convergent flow through Stainmore and

other terrestrial ice streams, although the proposed surge along the east coast of England (Catt, 1991b) may be invoked to explain the drumlins. The Wensleydale features could be explained by an unstable calving margin in Lake Humber which would have been in existence at this time in the Vale of York (Gaunt, 1981; Catt, 1991b). These interpretations clearly exemplify the critical role which the identification of former ice streams may have on the determination of glacial and deglacial events.

Nature of Ice Divide

An ice divide is an area of accumulation from which the ice will begin to flow outwards under gravity. Studies of the Antarctic ice sheets (Drewry, 1983) have shown that the general morphology of an ice divide is of a relatively narrow, linear zone composed of a series of ice domes which are separated from each other by ice saddles. Flowlines are lines of maximum surface slope and indicate that flow will generally diverge from ice domes and converge from ice saddles with the size of an ice dome reflecting glaciological stability (Hughes, 1981; Hughes, *et al.*, 1985). An ice sheet in equilibrium is thought to be dominated by a single central ice dome with the flowlines indicating radial flow to ice margins which are equidistant in all directions from the dome (Hughes, 1987), although such a model has been seriously questioned by other research workers (*cf* Andrews, 1987; Dyke and Prest, 1987; Boulton and Clark, 1990).

Drumlin distribution (Fig. 10) indicates that an ice divide was centred across the area from Wild Boar Fell (SD 7598) southwards across Baugh Fell (SD 7392) towards Great Coum (SD 7084) and lying across both Garsdale and Dentdale. From the pattern of ice convergence, the reconstruction indicates that domes and saddles can be identified along the line of ice shedding (Fig. 16). This interpretation of ice flow pattern shows that a restricted local ice dome over Baugh Fell, which is well established in the literature (*cf* King, 1976; Johnson, 1985; Catt, 1991a), is incorrect and that an ice divide developed over the Western Pennines, as a thin linear zone, 2 to 5 km in width over a distance of 15 km, and was independent of topography (Mitchell, 1991). Furthermore, this ice divide can be traced westwards across the Howgill Fells towards the Lake District (Letzer, 1978) indicating that it is not a local feature, but forms a major line of ice shedding in northern England (Mitchell, 1991).

Variations in ice flow direction, identified by the superimposed drumlins, reflect changes in location of the ice divide. Change from Ice flow event 1 to Ice flow

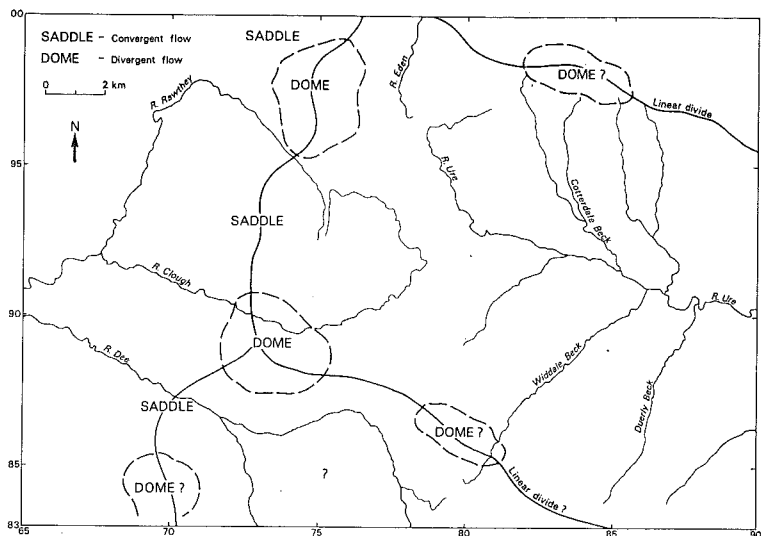


Figure 16. Nature of the ice divide in Western Pennines, showing possible location of former domes and saddles along its length (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

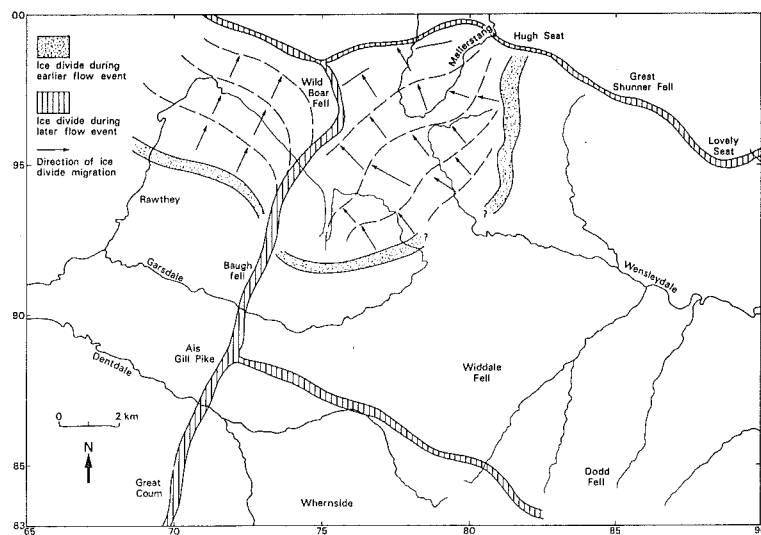


Figure 17. Possible directions of ice divide migration due to ice stream 'capture' (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

event 2 is demonstrated by an expansion of the ice divide northwards to include the high ground of Swarth Fell/Wild Boar Fell (Fig. 17) and which changed the relative importance of certain areas of the ice divide by instituting capture of source areas to both the Wensleydale and Rawthey ice masses at the expense of ice flowing northwards to the Vale of Eden (Fig. 17).

Palaeoglaciological Reconstructions

Variations in drumlin parameters across the study area have been shown to reflect changes within the complex dynamic system of the subglacial environment (Mitchell, 1991). This allows emphasis to be directed towards general statements reflecting the overall mechanical conditions of the ice and which are most easily seen in terms of driving stresses, as is the case with current analyses of present day ice sheets (cf Cooper, *et al.*, 1982; Drewry, 1983; Whillans, 1987).

The distribution of driving stresses is known for certain parts of the West Antarctic Ice Sheet with the lowest stresses (<20 kPa) at ice divides or where there are subglacial water bodies (Drewry, 1983). The highest driving stresses (>100 kPa) are found in areas of flow convergence at the head of outlet glaciers and ice

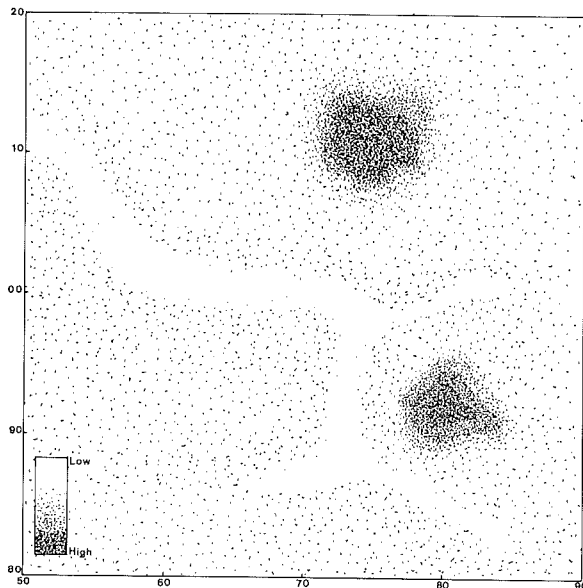


Figure 18. Suggested distribution of driving stresses under the Dimlington Stadial ice sheet in the area of the western Pennines and southern Vale of Eden (from Mitchell, 1991).

streams (Cooper, *et al.*, 1982; Whillans, 1987). This information can be used to tentatively reconstruct the basal stress field, within the Western Pennines and southern Vale of Eden (Letzer, 1978) which shows the highest driving stresses within the Wensleydale ice stream and towards Stainmore with low stresses in the vicinity of the ice divides (Fig. 18).

Areas of flow convergence will delimit the catchment area of an ice stream and will be the zone where the ice begins to accelerate (Hughes, 1981). It is therefore of note that at the head of the Wensleydale ice stream, which is where the flowlines converge and the linear ice stream commences, that the longest and most elongate drumlins occur (Mitchell, 1991). This suggests that high driving stresses/mass flux are accommodated by fast ice flow as the ice begins to stream (Fig. 15).

Deglaciation of the Dimlington Stadial Ice Sheet

Landforms associated with glaciofluvial activity, particularly meltwater channels and ice stagnation mounds have been mapped across the Western Pennines and southern Vale of Eden (Letzer, 1978; Mitchell, 1991). From these landforms, particularly the general absence of ice stagnation features, it is thought that the deglacial pattern was primarily one of ice backwasting to the ice divides, but with distinct differences between the western sector of the Rawthey valley, Garsdale and Dentdale, and the eastern sector dominated by Wensleydale. In the Western Pennines, there is some geomorphological evidence in the form of fragments of moraine ridges in the Rawthey valley and Dentdale, to suggest that the pattern of deglaciation involved temporary halts in the ice margin as the ice thinned and began to act as valley glaciers (Moulson, 1967; Mitchell, 1991). No such stillstands were identified in the Wensleydale sector, apart from in a small southern tributary, Bardale. It is suggested that deglaciation was rapid in Wensleydale, since there is no alteration to the pattern of drumlins across the area, but more step-like with stillstands and localised stagnation in the western valleys. The large meltwater channels indicate the direction of subglacial drainage. Since many of the large channels are cut through drumlins, they indicate later events in deglaciation.

Longitudinal landforms can be used to draw lines of ice retreat and reconstruct the pattern of ice wastage (Boulton, *et al.*, 1985). For the British ice sheet, this creates a number of concentric circles which are convex down ice reflecting the radial nature and divergence of flow during ice wastage in the lowland areas and defining the surrounding upland source areas to which the ice sheet is thought to

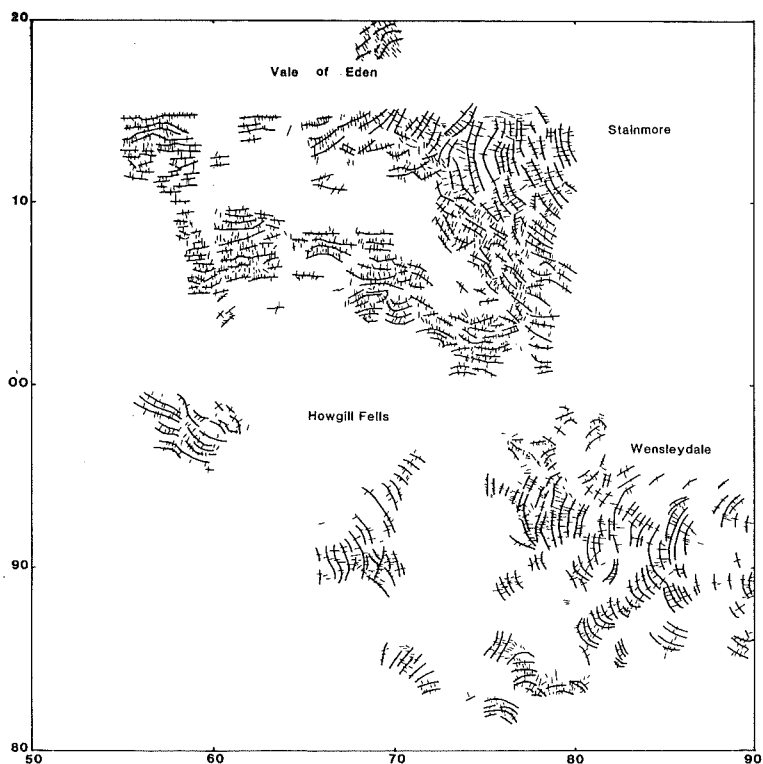


Figure 19. Pattern of ice recession determined for the western Pennines and the southern Vale of Eden based on the evidence from drumlins (data from Letzer, 1978, Whiteman, 1981, M. J. Brunskill, unpublished, Mitchell, 1991). Scale and orientation given by National Grid coordinates.

have retreated. However, whilst this has been done at the scale of the ice sheet, there have been no published attempts to reproduce the pattern of ice wastage for a smaller area for which the detailed longitudinal evidence is known. Using field evidence from individual drumlins and drift tails within the Western Pennines and southern Vale of Eden, lines of ice marginal retreat have been constructed at right angles to the long axes of the streamlined bedforms (Fig. 19).

In Wensleydale, the lines of ice retreat form a series of concentric circles which are concave down glacier indicating the overall pattern of ice flow of the upper sector of the former ice stream in the dale. The regularity of the lines and the lack of field

evidence to indicate breaks in the pattern of ice retreat, suggest that the ice stream remained active whilst the ice margin retreated. The ice sheet maintained a flow regime which was independent of topography throughout this period since there is no evidence is flow direction changing to divergence as individual ice fronts developed within the different dales except in a few small areas in the upper parts of tributary valleys, such as Widdale, Mossdale Moor and the flanks of Baugh Fell (Fig. 19). This suggests that only when the ice margin had retreated into the upper parts of these valleys, was the ice sheet thin enough for topographic control to influence ice flow direction.

The recessional pattern determined from the drumlins also shows a dominant flow convergence on Stainmore which agrees with the classification of this as an ice stream (Fig. 19). In most of the Vale of Eden, no distinctive trends are obvious,

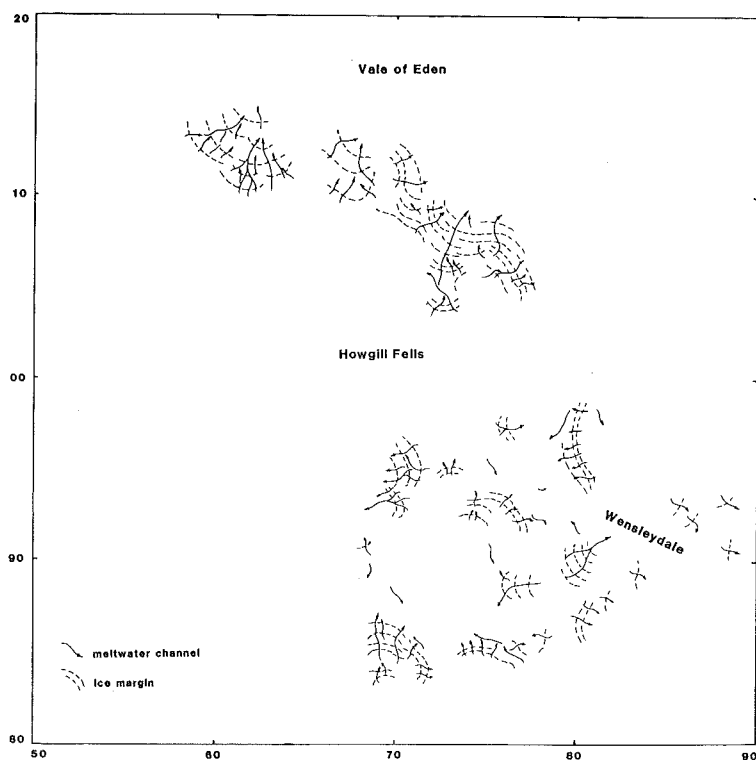


Figure 20. Pattern of ice recession determined for the western Pennines and the southern Vale of Eden based on the evidence from meltwater channels (data from Letzer, 1978, Mitchell, 1991). Scale and orientation given by National Grid coordinates.

and the drumlins indicate formation under active ice, thick enough to be independent of topography. Ice flow from the Howgills was not uniform along the east-west length of the Lune valley, but shows smaller concave patterns of ice flow, which suggest incipient streaming possibly associated with ice thinning and the beginning of a topographic influence, since these patterns are north of the major Howgill valleys. Slight convexity is seen at the entrance to some of the valleys in the northern Howgills, but this is very slight and even at the mouths of these valleys there is a convergence of flow into the main drumlin field. This evidence also indicates a period of flow from, and recession back to, the northern flanks of Wild Boar Fell and that the ice did not become topographically constrained as valley glaciers in this area.

A similar approach can be attempted by using the pattern of meltwater channels (Fig. 20), based on the principle that the channels were formed by meltwater directed to the ice margin by differences in hydrostatic pressure. In many areas such as the Rawthey valley and around Crosby Ravensworth, the meltwater channels indicate a convergent pattern with the smaller channels associated with secondary flow drawn to the main feeder channels which are eroded through the drumlin fields. However in other areas, there is a change from dendritic to radial drainage, such as in Dentdale, on Lunds Fell and on Smardale Fell which can be interpreted as reflecting that the meltwater channels were formed when the ice had receded so much that all that was left was a number of isolated plateau ice caps (Mitchell, 1991). This suggests that the meltwater channels reflect later events in deglaciation than the drumlins. This information also indicates that during deglaciation, the former ice divide became disjointed with saddles deglaciated first, since the ice will be thinner at these points with the final ice to exist at the location of former domes (Fig. 16).

Paraglacial Readjustment of Slopes to Deglaciation

The initiation of non-glacial geomorphic systems following deglaciation in an area which has been fundamentally altered by the presence of glacier ice, may mean that many geomorphic aspects of a mountain environment became potentially metastable, particularly in a periglacial situation (Barsch and Caine, 1984; Johnson, 1984; Harris, 1988). This response has been defined as one of 'paraglacial adjustment' (Church and Ryder, 1972) or as paraglacial 'conditions' and 'environments' (Johnson, 1984); one aspect of this complex environmental response in the Western Pennines was the failure of many slopes after deglaciation (Mitchell, 1991).

Attention was drawn to the slope failures for two reasons. Firstly, when mapping

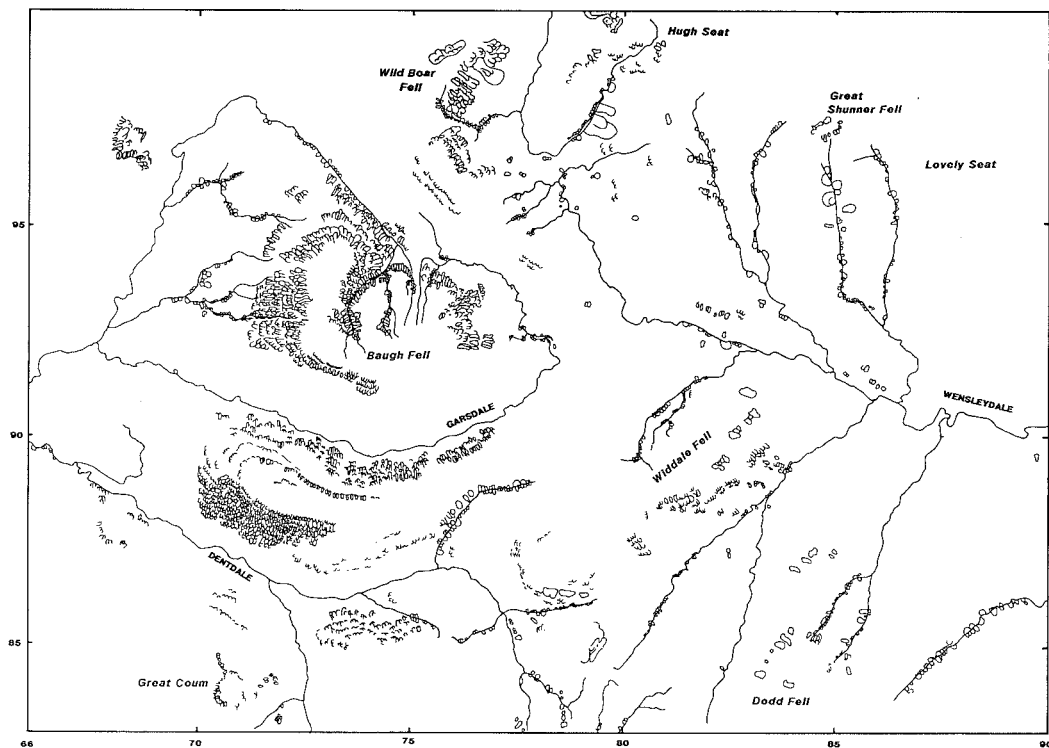


Figure 21. Distribution of mass movement in Western Pennines, showing major lobe forms. Enclosed polygons define more complex slope failures involving rock displacement (from Mitchell, 1991).

the drumlins in the bottom of the valleys, it became apparent that fluvial erosion had undercut the drumlin flanks and initiated small scale rotational failures and debris flows which are destroying the drumlin form. Secondly, mapping has shown that many of the hillslopes in the western dales are covered by massive and complex failures from large rock slope failures (RSF) to small active debris flows (Fig. 21). The high density of forms in the western dales is a direct response to the underlying geology where the upper parts of the fells are dominated by thick sequences of mudstones and shales.

This can be spectacularly demonstrated on the glacially eroded slopes of Mallerstang, where specific geological conditions, along the flanks of Wild Boar Fell, Little Fell, and the various scars on the eastern side of the valley have resulted in extensive and progressive slope failure (Fig. 21). The major failures are large mudslides and complex rotational deep seated landslides below the back scarp, with retrogressive debris flow activity and large boulder streams (Rose and Mitchell, 1989; Mitchell, this guide).

The age of these landforms is not yet established. However from their relationship to former recessional ice margins and meltwater channels, it can be demonstrated that many of these mass movements were first initiated as a paraglacial response to deglaciation of the Dimlington Stadial ice sheet and are of Lateglacial age (Mitchell, 1991). During the early stages of deglaciation, there would have been a great deal of water available at this time to trigger slope instability by increasing pore water pressure and reducing shear strength. The failures need not have happened immediately the ice left the area because there will be a lag time during which the stress field comes into a critical state for failure (Holmes, 1984; Ballantyne, 1986, 1991).

The occurrence of Loch Lomond Stadial moraine ridges and protalus ramparts lying on top and beneath landslide debris also indicates that slope instability occurred during and after this episode, probably during the early Holocene (cf Holmes, 1984; Ballantyne, 1986, 1990, 1991). Activity was probably encouraged by the return of cold climate conditions during the Loch Lomond Stadial. Once initiated, these mass movements would have continued intermittently over a period of time, probably well into the Holocene before achieving stability (Johnson, 1987).

LOCH LOMOND STADIAL LANDFORMS AND PALAEOGLACIOLOGICAL RECONSTRUCTION

W.A.Mitchell

The climatic deterioration of the Loch Lomond Stadial is marked in northern England by the re-establishment of small ice masses, particularly cirque and small valley glaciers, to about 70 locations in the high mountains in the Lake District (Sissons, 1980a) and in the Pennines (Manley, 1959; Gunson, 1966; Mitchell, 1991). In the Pennines, detailed mapping has only been completed in the Western Pennines where five small cirque glaciers, the largest being 0.4 sq. km in size (Table 3), became established during this period of climatic deterioration (Fig. 22). This can be compared to the size of glaciers which have been mapped in the Lake District where 76% of them were less than 0.5 sq. km in size (Sissons, 1980a) and to the much larger ice masses in Western Scotland showing that the Pennines were clearly marginal to glaciation at this time (Gray and Coxon, 1991).

- Cirque glaciers in the Western Pennines are recorded at Cautley Craggs, Combe Scar, Great Coum, Swarth Fell and Whernside (Fig. 22) (cf Site reports for Cautley Craggs, Combe Scar and Swarth Fell in this guide). All five glaciers identified and mapped in the area are defined in their lower reaches by terminal moraine ridges. A number of ridges in the Mallerstang have also been interpreted in the literature as lateral and terminal moraine ridges of eight small cirque glaciers (Rowell and Turner, 1952), although their glacial origin was questioned by King (1976) who explained them as periglacial features and Letzer (1978) who proposed that they were either rock glaciers or rotational landslides. These have recently been mapped at 1:3,000, and apart from the moraine ridge at Swarth Fell, can all be more simply explained by reference to features associated with slope failures (Table 4), particularly mudslides and deep seated rotational failures (Mitchell, 1991; this guide).

Glacial Reconstructions

Reconstructions of former Loch Lomond Stadial palaeoglaciers have been completed for a number of different areas of the British Isles (Sissons, 1974b, 1977a,

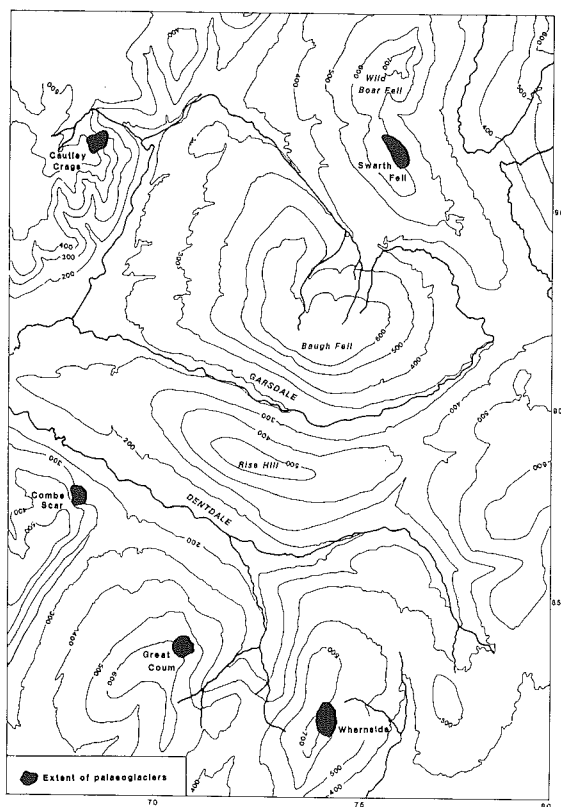


Figure 22. Location of Loch Lomond Stadal glaciers in the Western Pennines (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

Paleoglacier	Altitudinal Range (m ²)	Glacier Area (km ²)	Max. Ice Thickness (m)	Mean Ice Thickness (m)	Mean Ice Volume (km ³)
CAUTLEY CRAGS	330 - 600 (270)	0.24	75	26.6±21.9	0.007
COMBE SCAR	260 - 422 (218)	0.22	60	24.2±17.4	0.006
GREAT COUM	512 - 610 (98)	0.21	39	22.5±9.5	0.005
SWARTH FELL	520 - 632	0.37	34	14.6±9.0	0.005
WHERNSIDE	578 - 700	0.40	41	31.8±9.8	0.013
Lake District Summary (64 Glaciers) (Sissons, 1980b)		49 glaciers < 0.5 km ² (Total Area covered 54.55 km ²)	240	60	52 glaciers <0.1 km ³

Table 3. Dimensions of Loch Lomond Stadal palaeoglaciers in Western Pennines (from Mitchell, 1991).

Geomorphic Evidence	Interpretation of Rowell and Turner (1952)	Interpretation in this guide
topography	mounds and hollows, boulders in great 'festoons' suggestive of moraines	depositional zone of landslides as lobes
cirques (corries)	hollows in hillside on the different mountains formed by glacial erosion	source areas (scars) of large rotational failures, to produce a scalloped backwall
transverse moraines	lower limit of small glaciers	compression ridges in depositional zone of mass movement
lateral moraines	sharp lateral limit of glacier	levees of debris flows and mudslides
corrie lakelets	lakes impounded behind moraine ridges	drainage impeded by rotated blocks of bedrock
composition	boulders of grit moved by a glacier	boulders carried on surface of debris flows and mudslides

Table 4. Different Interpretations of the geomorphological evidence in Mallerstang with respect to the identification of former glaciers (from Mitchell, 1991).

1977b, 1979a, 1980a, 1980b; Sissons and Sutherland, 1976; Cornish, 1981; Gray, 1982; Sutherland, 1984; Ballantyne, 1989, 1990). Similar research has been completed for comparable areas of former mountain glaciation in North America (Porter, 1977; Pierce, 1979; Meierding, 1982; Leonard, 1984, 1989; Murray and Locke, 1989; Locke, 1990). This has involved the delimitation of the former glacier from the geomorphic evidence, and contouring of the glacier

surfaces to allow the overall dimensions of the former ice masses to be determined (Table 3). Such palaeoglacier reconstructions have been used in two distinct but related ways. Firstly, they have been used in the interpolation of regional climatic conditions for the period of geological time that the glaciers were in existence by calculating the equilibrium line altitude (ELA) (Sutherland 1984; Locke, 1990). Secondly, they can also be used to determine the palaeoglaciology of the former ice masses by reconstructing the basal shear stress distribution and flow characteristics (Pierce, 1979; Porter, *et al.*, 1983; Murray and Locke, 1989; Sharp, *et al.*, 1989).

The procedures which underlie these reconstructions are seldom discussed in the literature. However it must be appreciated that there is a basic assumption that the glaciers have achieved mass balance equilibrium (Sutherland, 1984; Leonard, 1984; Murray and Locke, 1989). It is also important to appreciate the critical nature of the contouring of the former ice surface since very small increases in contour curvature can lead to significant changes in ice thickness, particularly on small ice masses (Mitchell, 1991).

Palaeoclimatic Reconstructions

The parameter which has been most commonly used to allow palaeoclimatic inferences is the firn line or equilibrium line altitude (ELA) (*cf* Locke, 1990). The ELA is regarded as the critical parameter in climatic reconstructions since studies of present day glaciers have shown that there is a positive correlation between the ELA and climate (Osmaston, 1975, 1989; Sutherland, 1984; Ballantyne, 1989; Leonard, 1984, 1989; Locke, 1989, 1990).

In reconstructions of Loch Lomond Stadial glaciers in Great Britain, the ELA has been calculated from the area-weighted mean altitude (AWMA) (Sissons, 1974b; Sissons and Sutherland, 1976; Sutherland, 1984; Ballantyne, 1989). This assumes that the accumulation and ablation gradients are equal and gives an ELA which is related to palaeoclimatic conditions which occurred when the former glacier was at its maximum extent and in steady state (Sissons, 1974b; Sutherland, 1984).

However, there are a number of other methods which have been used to find the altitude of the ELA. These include the upper altitude of lateral moraines (MOR), the lowest elevation of the cirque floor (CIR), the toe-headwall altitude ratio (THAR) (Meierding, 1982; Sutherland, 1984; Pierce, 1979; Murray and Locke, 1989;

Locke, 1990) and the accumulation-area ratio (AAR) with a common percentage of accumulation area to total glacier area being 65% (Porter, 1975; Zwick, 1980; Meierding, 1982; Leonard, 1984; Murray and Locke, 1989).

Each of these methods will give a different value for the ELA of the glaciers in the study area (Table 5)(cf Meierding, 1982). Mean values show that the values of ELA range from a CIR with a value of 456 ± 138 m to Highest Lateral Moraine at 558 ± 68 m. A THAR (40%) ELA has a mean value of 506 ± 114 m and an AAR (65%) ELA has a mean value of 485 ± 129 m. Most of the published studies from North America indicate that an AAR (65%) gives the most accurate results within \pm c. 50m of the original ELA (Meierding, 1982; Leonard, 1984; Murray and Locke, 1989). The AWMA technique gives a mean ELA of 502 ± 104 m. Mean values for all the different techniques have been calculated and show close similarity to each other with quite low values of standard deviation (Table 5).

Because of the various uncertainties of the methods, a mean value from AAR (65%) and AWMA techniques has been calculated and used in the analysis of the Western Pennine glaciers (Mitchell, 1991), giving a mean ELA of 494 ± 122 m with a range from 311m to 608m, although the small size of the sample must be noted. This compares with the mean ELA altitude of 527m (range 284-785m) for the 62 glaciers mapped in the Lake District (Sissons, 1980a). Within the study area, the lowest values occur in the two western cirques of Cautley Craggs (428m) and Combe Scar (311m). The ELA values of the other three glaciers are distinctively higher, ranging from 555m (Great Coum) to Swarth Fell (567m) with the highest being Whernside with an ELA of 608m.

A contour plot of the ELA distribution in the Western Pennines shows that there is a trend across the area with the lowest ELA values occurring along the western flanks of the upland area of the Askrigg Block (Fig. 23). This information has been used to infer the direction of snow bearing winds which would be blowing from the direction of lowest to highest ELA (Sissons, 1979a; Sissons and Sutherland, 1976; Ballantyne, 1989) although it has also been demonstrated that glacier ELA where snowblow is an influence may reflect both southerly snow bearing and southwesterly snow blowing winds (Ballantyne, 1989). In this area the ELA trend would suggest that snowfall is associated with westerly air streams (Mitchell, 1991). There is no indication that a more southerly flow of snowbearing winds was important in the Western Pennines, since if such winds had been important, glaciers would have developed along the southern margin of the Askrigg Block.

	CAUTLEY CRAGS	COMBE SCAR ^a	GREAT COUM ^a	SWARTH FELL	WHERN'SIDE	<u>x</u>	
A LOWEST CIRQUE ¹ (CIR)	350	270	520	560	580	456	±138
B THAR ² 35%	425	333	552	562	622	498	±117
40%	439	343	556	568	626	506	±114
45%	453	353	561	575	635	515	±112
C HIGHEST LATERAL MORAINES ³	460	none	560	600	610	558	±68
D AAR ⁴ 70%	411	285	543	552	604	479	±129
65%	418	292	548	558	609	485	±129
60%	427	299	552	564	613	491	±127
E AREA WEIGHTED ⁵ MEAN ALTITUDE (AWMA)	437	329	562	575	607	502	±104
F MEAN VALUE (all methods)	424±32	313±30	550±13	568±14	612±16	493	±123
G MEAN VALUE (AAR 65% and AWMA)	428±13	311±26	555±10	567±12	608±1	494	±122

1. Lowest altitude of cirque floor.

2. Toe - headwall altitude ratio.

3. Altitude of lateral moraine.

4. Accumulation Area Ratio.

5. AWMA $f = \frac{A_i h_i}{h}$

a - values derived from second reconstruction of ice surface.

Table 5. Values of ELA determined by five different methods (A-E) with mean value for all methods (F) and mean for AAR and AWMA methods (G) (from Mitchell, 1991).

It may well be the case that local environmental conditions rather than regional depression of the snowline are responsible for the low ELA values particularly in the western palaeoglaciers. It has been shown that additional mass to the glaciers by snowblow is a critical factor in determining the ELA of the glacier in other mountain areas of the British Isles by defining snowblow areas which are adjacent and upslope of the former accumulation areas (Sissons, 1979a, 1980a, 1980b; Sutherland, 1984; Ballantyne, 1989). The problem of mass being added to the glacier by upslope transfer of snow has been investigated for South Wales palaeoglaciers where an arbitrary limiting gradient of 9° was used to define snowblow areas (Robertson, 1988).

In the Western Pennines potential snowblow areas have been defined as including ground above the ELA which is laterally continuous to the glacier (Mitchell, 1991). Potential snowblow areas have been calculated for each 15° sector for each of the five glaciers in the western Pennines (Fig. 24), and this has been used to determine

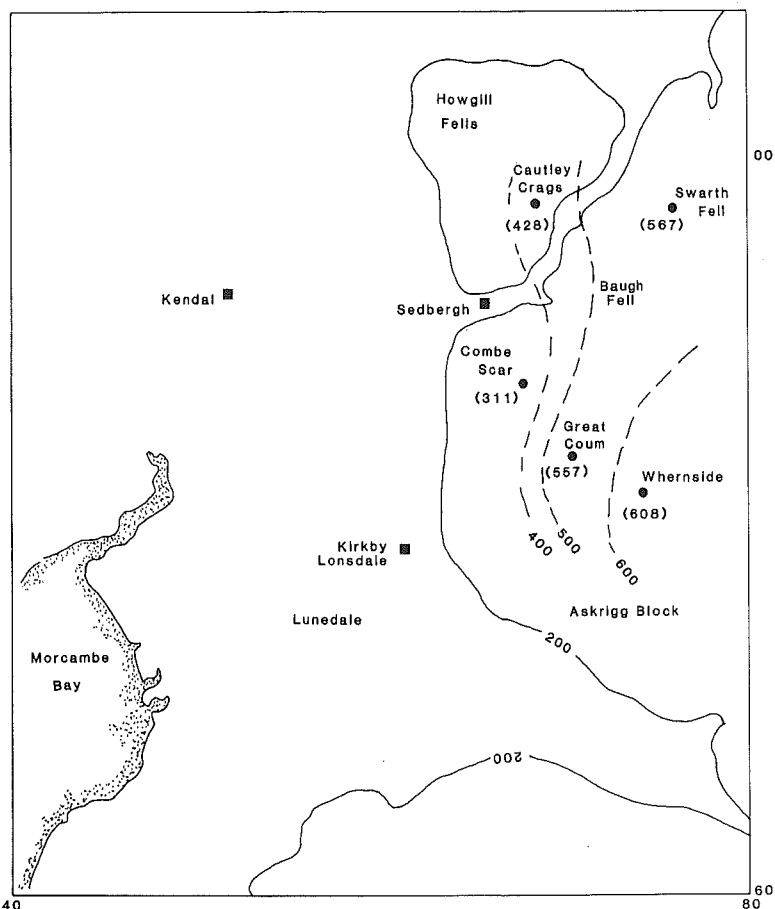


Figure 23. Contour lines of ELA values for the Western Pennines (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

the snowblow factor (cf Sissons, 1980a) (Table 6).

A significant negative correlation has been found between ELA total snowblow factor which indicates the control of snowblow in lowering ELA values. For the different 90° sectors, only the western sector (226-315°) gives a significant negative correlation which suggests that glacier development and sustenance was controlled by snow accumulation associated with westerly air streams. Although there is no significant correlation with the southwest sector values, the high snow blow factor values achieved by Cautley Craggs and Combe Scar glaciers for this

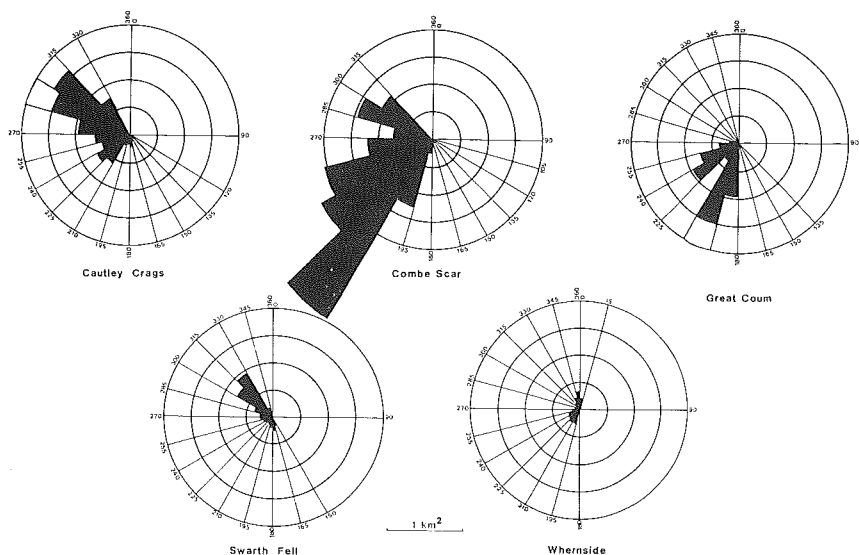


Figure 24. Polar plots of potential snowblow area for each glacier (from Mitchell, 1991).

GLACIER	GLACIER AREA (km ²)	ELA (m)	iF	SNOWBLOW FACTOR (90° SECTORS)						MEAN SNOW BLOW FACTOR FOR GLACIER	RATIO OF SNOWBLOW AREA TO GLACIER AREA
				NE	SE	S	SW	W	NW		
Cautley Crags	0.24	428	11.5	0	1.20	2.29	3.44	5.03	4.53	5.81	33.81
Combe Scar	0.23	311	15.0	0	1.13	4.94	6.91	6.22	3.79	7.95	63.30
Great Courn	0.21	555	17.5	0	0.54	3.70	4.82	3.38	1.33	5.03	25.29
Swarth Fell	0.37	567	16.5	0	1.02	1.52	1.64	2.23	2.56	3.20	10.29
Whernside	0.40	608	15.1	0.66	0	1.23	1.73	1.39	0.82	2.34	5.49

Table 6. Snowblow Factor values for the palaeoglaciers (from Mitchell, 1991).

orientation, suggest that southwesterly winds will also have played a role in snowfall and glacier growth, as suggested by Payne and Sugden (1990).

Palaeoglaciological Reconstructions

Knowledge gained from theoretical work in glaciology and the study of present day ice masses (Schilling and Hollin, 1981) has been recently used in conjunction with glacial geological evidence to determine the glaciology of former ice masses, in terms of basal shear stress, glacier velocity and mass balance (Pierce, 1979; Murray and Locke, 1989; Sharp, *et al.*, 1989). Basal shear stress, or driving stress, is an important control on the nature of ice motion with respect to nature of internal deformation or basal sliding. These calculations can be determined from the same glacier reconstructions used to establish the ELA (Mitchell, 1991).

Using the reconstructed ice surface contours, ice thickness and surface slope can then be determined and placed into the standard equation for basal shear stress for a valley glacier (*cf* Paterson, 1981; p. 103);

$$\tau = \rho g H F \sin \alpha$$

where τ = basal shear stress, ρ = specific gravity of ice, g = gravitational acceleration, H = ice thickness, F = shape factor to allow for lateral drag on valley sides (Nye, 1965), and α = angle of surface slope.

This is a modified form of the general basal shear stress equation since it takes into consideration, F which is defined by the effects of lateral drag along the valley sides (Nye, 1965; Pierce, 1979; Murray and Locke, 1989). F is determined by calculation of half glacier width divided by mean ice thickness (Pierce, 1979).

Basal shear stress has been calculated along a central long profile line for each glacier using a sample distance of 50m (Fig. 25). The mean stress values show a range from 0.35 to 1.04 bars, within the range of present day small glaciers (Paterson, 1981; Murray and Locke, 1990). It is only with the western palaeoglaciers that high values (>1.5 bars) are obtained where the ice was at its thickest and on relatively steeply sloping ground, thereby increasing surface slope. These values can then be used to define glacier flow regime with extending flow with high values and compressive flow with lower stress values (*cf* Pierce, 1979).

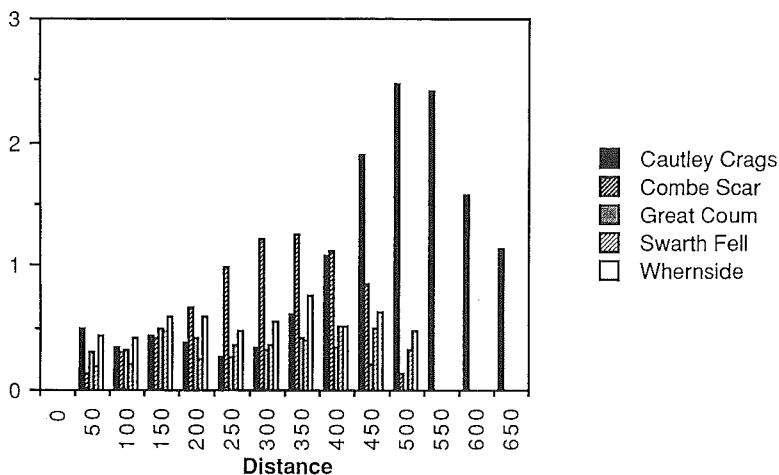


Figure 25. Variations in basal shear stress along a central longitudinal profile for each glacier reconstruction. Distance is measured from the glacier terminus (from Mitchell, 1991).

Ice velocities can also be calculated using standard equations (cf Murray and Locke, 1989; Sharp, *et al.*, 1989), such that:

$$V_c = 2A\tau_b^n H / (n+1)$$

where, V_c is the central line surface velocity due to ice deformation and is defined as the difference between surface velocity (μ_s) and basal velocity (μ_b), where μ_b is zero; A is the constant in Glen's Law, H is ice thickness and n is the exponential of the flow law (≈ 3); τ_b = basal shear stress.

There is a problem in reconciling the glaciological reconstructions and the large constructional landforms found at the ice margins of the five palaeoglaciers, since the basal shear stress values and the corresponding ice velocities are too small to have formed the terminal moraine ridges (Table 7).

Previous work on other Loch Lomond Stadial palaeoglaciers in the British Isles has interpreted the large terminal moraine ridges as indicating glacier equilibrium,

	Ice Thickness		Steady State	Advance	% increase
	Steady State	Advance Mode	V_c (ma^{-1})	V_c (ma^{-1})	
CAUTLEY CRAGS	23	42	0.44 (3)	2.01 (3)	457
COMBE SCAR	44	55	3.46 (1)	8.76 (1)	253
GREAT COMBE	30	47	0.21 (4)	1.08 (5)	540
SWARTH FELL	25	42	0.10 (5)	1.10 (4)	1100
WHERNSIDE	35	53	0.57 (2)	3.45 (2)	605

Table 7. Calculated ice velocities for the reconstructed glaciers showing the mean ice thicknesses and surface velocity for steady state (mass balance) mode and advance mode; percentage increase relates to the recalculations for velocity in advance mode. Figures in brackets show rank order for the palaeoglaciers. (from Mitchell, 1991).

with the ice margin stationery for a long period of time, that is size of moraine reflects the length of time at ice margin (Sissons, 1980a; p.26). However, the palaeoglaciological reconstructions indicate that steady state conditions are not appropriate for understanding the geomorphological response of the glacier and that moraine construction requires advancing glaciers. This has been explained by reconstructing the glaciers in an advance mode which increases the ice thickness, particularly in the lower parts of the glacier and gives the necessary increase in shear stress and velocity to push the material into ridges.

WIDDALE

W.A.Mitchell

Introduction

Widdale (SD 8388) is one of the main southern tributary valleys of Wensleydale (Fig. 1), extending from the watershed between Wensleydale and Ribblesdale at Newby Head Moss (SD 795838), northeast to the confluence of Widdale Beck with the River Ure at Appersett (SD 859907). To the west, the ground rises to the plateau of Widdale Fell (SD 789871) with a summit at Great Knoutberry Hill (672m). To the east, Snaizeholme Fell (540m) separates Widdale from its tributary valley of Snaizeholme (SD 8386), which is one of the few valleys which can be defined as a glacial trough (Clayton, 1966). To the east of Snaizeholme, Dodd Fell (SD 841845) forms a plateau, which becomes Ten End north towards Wensleydale.

The landforms found in Widdale are typical of this part of the dales, and have been visited to demonstrate geomorphological mapping technique, the nature of the drumlins in terms of morphological variability and post-depositional modification by fluvial erosion and mass movement, and to examine an exposure in a drumlin at Widdale Side.

Geomorphological Mapping

This dale is a good area to demonstrate the technique of field mapping and point out the problems of defining simple geometric forms, such as drumlins where they have been subjected to post-depositional erosion and modification. This area was mapped at an early stage of a research project (Mitchell, 1991), when attention was focussed on the glacial landforms and which led to the definition of the present valley of Widdale Beck as a meltwater channel (Fig. 26a). As mapping proceeded within the dales to the west of Widdale, the significance of mass movement features became appreciated, and the area was re-mapped (Fig. 26b) to illustrate the extent of post-depositional modifications to drumlins in valley bottoms in upland areas. Landslides are an important geomorphological element in these upland valleys, and must be properly delimited to allow the accurate reconstruction of the

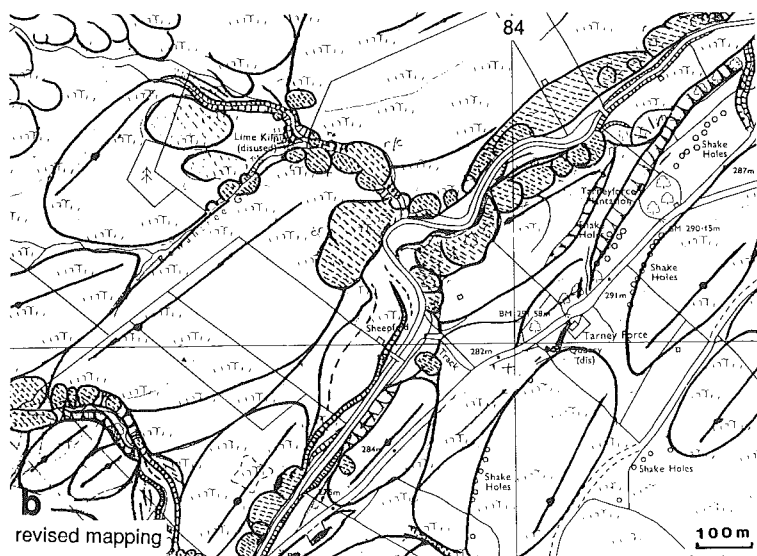
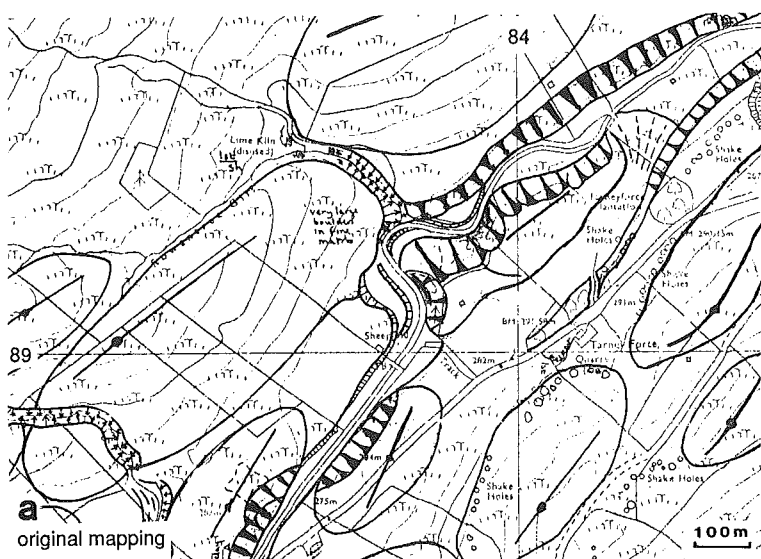


Figure 26. Comparison of original field mapping (a) and revised mapping (b) to take into consideration the mass movements on drumlin flanks in valley bottoms; example from Widdale. Original mapping scale 1:10,000 (from Mitchell, 1991). (Crown copyright reserved)

original drumlin shape for meaningful morphometric analysis.

Drumlin Distribution

Widdale shows an extensive development of drumlins through much of its length from Widdale Head (SD 8085) and Mid Widdale (SD 8186), northeast to the confluence of Widdale Beck with the River Ure (Fig. 27). At Widdale Side (SD 8388), the drumlins are continuous with those in Snaizeholme. 111 drumlins have been mapped within the catchment area of Widdale Beck, including those within the lower part of Snaizeholme and on the summit plateau of Widdale Fell (Table 8). In the valley, there is no distinct break eastwards where the streamlined landforms continue as individual drumlins or drift tails across the end of Ten End (SD 8487). To the north the drumlins continue until just south of Bluebridge (SD 853905) where there is an area of mounds. Around Appersett, they occur south of a large drift tail which extends northeast from Widdale Fell and which is taken to define the northern margin of the Widdale drumlin sub-group.

The main group of drumlins lies within the valley bottom extending up the lower slopes (Fig. 27). They reach their highest altitude in the upper part of the valley, at Ling Beck, towards Cross Gate (SD 801865) and towards Swineley Pasture (SD 803858), where they achieve heights of around 500m OD with an altitudinal range is 192-508m OD (mean value of 330.1m OD). There is a full range of drumlin sizes within this group (Table 8) with small drumlins next to megadrumlins, particularly around Widdale Side (SD 830885) (Fig. 27). General trend of the drumlins is down the valley, with a range from 001° to 351° but with a mean of 57° in the southern part of the valley which gradually increases to 94° in the northern part.

A number of isolated drumlins have also been mapped in Snaizeholme. Their dimensions can be compared to those of Widdale (Table 8) and show a mean orientation of 45° which is oblique to the valley axis indicating an ice source area in upper Widdale, across Snaizeholme Fell which was relatively more important than the accumulation area of Grove Head to the south of the valley.

A further group of thirteen drumlins is distinguished by reason of altitude, occurring on the plateau top of Widdale Fell (Fig. 28; Table 8). They occur south of Widdale Great Tarn (SD 793877), and to the northeast of Little Knoutberry Hill

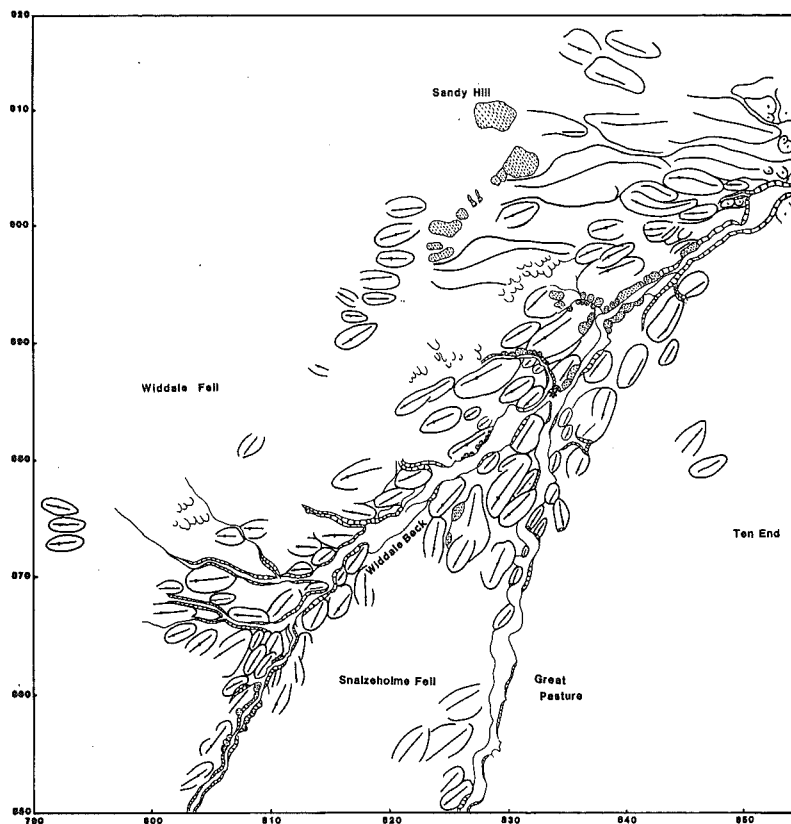


Figure 27. Geomorphological map of drumlin distribution in Widdale and Widdale Fell (from Mitchell, 1991). Scale and orientation given by National Grid coordinates. * indicates location of exposure to be visited at Widdale Side (cf Rose, this guide).

(SD 811889) near Sandy Hill (SD 825904). Drumlin altitude varies between 562 and 651m OD with a mean of 590m OD, with a mean orientation value of 87° indicating easterly flowing ice from an ice source in Dentdale.

This information clearly indicates that former ice flow direction within Widdale was down valley, from a source area/ice divide along the present watershed between Wensleydale and Ribblesdale. The drumlins on Widdale Fell, and orientation within Snatzeholme both indicate that the ice sheet was thick enough to have completely inundated the present topography. Furthermore the ice crossing Widdale Fell indicates a former source area to the west between Dentdale and Garsdale where there

		ORIENTATION ($^{\circ}$)	LENGTH (m)	WIDTH (m)	AREA ($\times 10^{-4}$ sq km)	ER dimensionless	K dimensionless	Sd dimensionless
SNAIZEHOLME (8)	\bar{x}	44.8	391.4	165.6	511.8	2.56	2.68	0.32
	range	25-62	250-530	90-290	155-1044	1.69-3.59	2.02-3.56	0.17-0.5
WIDDALE (98)	\bar{x}	57.2	333.6	147.6	400.5	2.31	2.53	0.37
	range	001-351	90-805	40-400	29-1912	1.22-3.92	1.41-4.39	0.11-0.63
WIDDALE FELL (13)	\bar{x}	87.2	291.9	126.5	283.5	2.33	2.67	0.36
	range	64-121	225-430	100-170	80-504	1.64-3.46	1.49-5.70	0.25-0.53

Table 8. Summary descriptive statistics (mean and range) for the drumlins mapped in Widdale, Snaizeholme and Widdale Fell (from Mitchell, 1991).

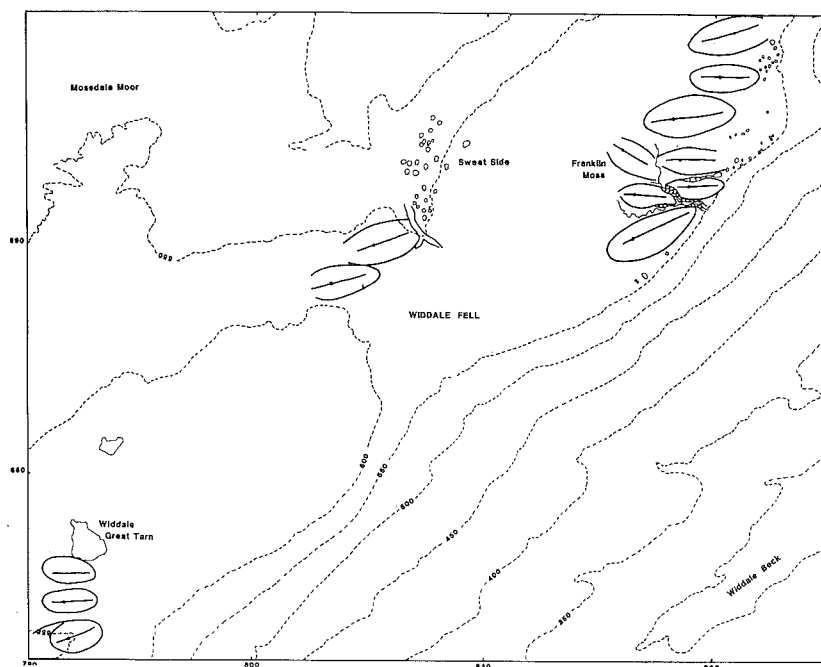


Figure 28. Drumlins on the plateau of Widdale Fell; their location relative to the drumlins in Widdale can be examined on Fig. 27. Scale and orientation given by National Grid coordinates. (from Mitchell, 1991).

must have been an ice divide (Mitchell, 1991). There is no evidence in this particular area to indicate that the ice source area was Baugh Fell/Wild Boar Fell, as previously envisaged (Goodchild, 1875; Dakyn, et al., 1891; Raistrick, 1926) and there is no evidence for ice flow up valley, particularly in Snaizholme, as previously interpreted (Raistrick, 1926; Clayton, 1966). The evidence clearly reflects an ice source to the south with ice flowing north to join the main Wensleydale ice.

DRUMLIN SEDIMENTS, WIDDALE SIDE

J. Rose

Introduction

Although an exposure near Askrigg in Wensleydale provided the location for a classical study of till fabric by Andrews and King (1968), there are, at the present time, very few exposures of *in situ* Quaternary sediments within the upper part of the western Pennines and of those that do exist, very few can be shown to relate clearly to the form of the glacier bedforms. One site where Quaternary sediments are currently well exposed within a well defined drumlin is at the confluence of Widdale and Snaizelhome Becks in Widdale (SD 834886). This site has been the subject of preliminary investigation with undergraduate students from Birkbeck College and Royal Holloway and Bedford New College, University of London, and some results of this work are described below.

Site Location

The exposure is located in the stoss end of a simple drumlin, located at the valley bottom of Widdale (Fig. 27). The exposure intersects the crestline of the landform which trends 045-225°, although the larger part is on the eastern side of the drumlin. The area exposed extends about 18m along the base with a maximum height of about 7m; analysis has been carried out at a number of sample points within each lithofacies (Fig. 29).

Lithological units

Four lithological units can be observed and some of the sedimentary properties are summarised in Table 9. These results were derived during a fieldwork training exercises and must therefore be used with caution as they are subject to relatively high operator variance.

The lowest unit (Unit 1) in the succession reaches a maximum observed thickness of c. 5m. It is a strongly overconsolidated very dark grey, matrix-dominated diamicton with clasts extending up to large boulder size. The matrix is composed of crushed mudstone, and incorporates many clasts which are mainly

Lithological Unit	Number of Sample Points	Matrix Munsell Colour	Lst.	Clast Sst.	Lithology Ist.	Mst.	Clast Ang.	Roundedness S.A.	S.R.	V.M. %age n = 50	Clast Signif.	Fabric R.V. Azim.	M.D.	Site No.
					(% values)			(Powers' classes) (% values)						
Unit 4														
Dark brown diamicton	4	7.5YR4/4	n = >200					n = >200		33.8	99.9	37°	24°	15
		-10YR3/3	51.0	46.0	3.0	0.0	20.3	41.7	38.0	27.0	99.7	58°	21°	14
										29.0	99.8	28°	24°	13
										18.9	n.s.		27°	12
Unit 3														
Upper very dark grey diamicton	2	5Y3/1 - 10YR3/1	n = >100					n = >100		29.0	99.8	86°	27°	11
			68.0	30.5	1.0	0.5	34.0	56.0	10.0					
Unit 2														
Silty sand with fine gravel	1	10YR3/1	n = 100					n = 100						10
			66.0	32.0	1.0	1.0	8.0	61.0	31.0					
Unit 1														
Lower very dark grey diamicton	9	5Y3/1 - 10YR2/1	n = >350					n = >200		28.4	99.8	75°	21°	9
			76.4	23.3	0.2	0.1	30.0	48.3	21.7	27.0	99.8	41°	26°	8
										20.0	n.s.		26°	7
										40.3	99.9	58°	21°	6
										55.0	99.9	18°	24°	5
										80.0	99.9	18°	17°	4
										33.7	99.9	28°	28°	3
										46.7	99.9	67°	27°	2
										19.9	n.s.		42°	1

Key: Lst. = limestone, Sst. = sandstone, Ist. = ironstone, Mst. = mudstone.

Ang. = angular, S.A. = sub angular, S.R. = sub rounded.

V.M. = vector magnitude; Signif. = significance; R.V. = resultant vector; MD = mean dip; n.s. = not significant.

Site No. = site location on Figure 29.

Table 9 Sedimentary properties of drumlin lithofacies at Widdale Side, North Yorkshire.

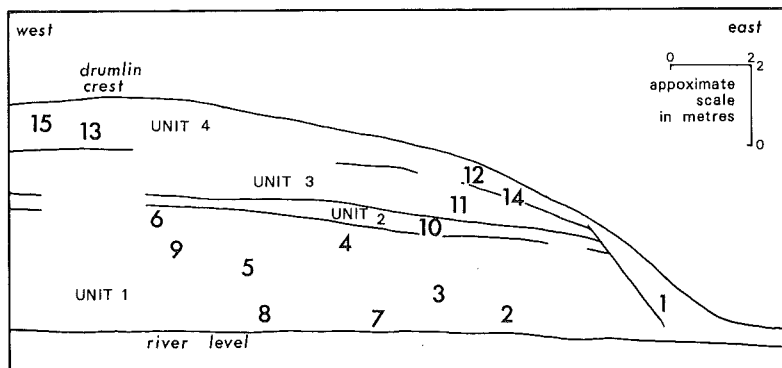


Figure 29. Schematic representation of sedimentary units exposed in the stoss end of a drumlin at Widdale Side, North Yorkshire (SD 834886). Numbers refer to location of sample points referred to in Table 9.

limestone. Together these lithological types form the most abundant bed material, but there are also a small number of sandstone clasts. Clast shape is predominantly sub-angular and angular, and many of the larger particles show a classical flatiron shape (Boulton, 1978). Striations are conspicuous on clasts of all sizes and show a trend closely parallel with that of the valley axis and drumlin crestline. Macrofabric analysis at 9 locations in this unit (Fig. 29) shows that in most cases, the elongate particles have a preferred orientation with $+30^{\circ}$ - -27° of the drumlin axis (Table 9). All properties suggest that this is a basal till unit formed of local bed materials and heavily stressed during lodgement by ice moving down Widdale.

One sample from Unit 1 shows a low consolidation, low fabric strength and high dip values. This site is located at the eastern margin of the exposure and is interpreted as sub-aerially slumped material, formed by mass movement caused by present day fluvial erosion.

Unit 2 is a continuous bed of silty sand, sand, and sand and fine gravel which reaches a maximum thickness of 0.6m. Within the exposure, this unit acts as an aquifer and promotes local slope failure. The fact that slope failures are developed around the drumlin (cf Fig. 26b) suggests that this unit is extensive throughout the bedform. The lithological composition of the clasts is similar to that of the overlying and underlying diamictons although it is more typical of overlying Unit 3

because of a slightly higher proportion of sandstone (Table 9).

Sorting properties, grain size distributions and the more highly rounded clast shapes indicate fluvial transportation and sedimentation, associated with flow conditions ranging from low energy silt deposition to relatively higher energy responsible for the transport and deposition of the medium gravel. There are no structures suggestive of ice contact, but the position of the unit within the drumlin and above the present valley bottom, indicates deposition either at a former valley bottom location which has since been eroded, river deposition within a confined position such as subglacially, or within ice walls.

Unit 3 is a moderately consolidated, matrix dominated diamicton with clasts extending up to boulder size. Relative to the lower very dark grey diamicton, this unit has a higher proportion of sandstone clasts and a higher level of clast angularity (Table 9). Only one macrofabric result is available, with a preferred orientation $+36^\circ$ from the valley axis/drumlin crestline. Although evidence is limited, information available suggests that only moderate stresses were associated with the deposition of this unit, which may be explained by subglacial melt-out processes.

Unit 4 forms a veneer, up to 1.5m thick, parallel with the surface of the drumlin. It is a loosely consolidated, dark brown, matrix dominated diamicton. Relative to Units 1 and 3, it has a high sand fraction, high sandstone clast content and higher proportion of rounded clasts (Table 9). Of four clast fabrics measured in this unit, three show a preferred orientation with a trend $+13^\circ/-17^\circ$ of the valley axis/drumlin crestline.

Some properties of this unit, such as the matrix colour and clast roundness could be explained by postglacial pedogenesis, but the higher proportion of sandstone clasts suggests derivation from a different source to the lower diamicton units (there is no evidence for 'ghosts' of limestone pebbles in this unit, which would lead to an increase in sandstone by removal of the soluble limestone). Sandstones are more abundant on the higher valley side slopes, whereas limestone and mudstone are more abundant in the valley bottom.

It is suggested, therefore that this upper unit is composed predominantly of supraglacial debris, forming a veneer of morainic drift across the surface of the landform (Eyles, 1979). The strongly preferred clast fabrics are interpreted as the

product of debris flow processes that contributed to the deposition of this supraglacial unit.

Overall Interpretation

Despite the amount of work done so far, evidence for the relationship between the deposits and the drumlin landform is far from clear. The four lithofacies reflect four different depositional environments, all of which appear to be related to glacial processes: Units 1 and 3 were formed by direct glacial deposition, Unit 4 by debris flow across the glacier surface and Unit 2 by glacial meltwater. Stress fields in Units 1 and 4 closely parallel the trend of the drumlin but the evidence is derived from a 2-dimensional exposure and there is insufficient evidence to say whether this relationship is coincidental (as is assumed for the supraglacial morainic drift of Unit 4) or causal (for Unit 1) (see Rose, 1989a for three dimensional analysis of sediments forming a glacial bedform).

For the purposes of stimulating discussion, the following explanation is put forward to explain the history of deposition and the relationship of these deposits to the drumlin landform.

- a) Unit 1 formed as the main body of the drumlin bedform by lodgement.
- b) Unit 2 formed subglacially by a meltwater river with a variable discharge in a cavity between the ice and the drumlin.
- c) Units 3 and 4 formed respectively by subglacial meltout and supraglacial debris flows as a carapace around the main body of the drumlin.

SWARTH FELL

W.A.Mitchell

Introduction

This mountain forms the high ground, reaching a height of 681m OD, between the upper part of Grisedale and Ais Gill Moor in upper Wensleydale (SD 7596)(Fig. 30) and is joined to Wild Boar Fell which forms a higher mountain area to the north. Along the eastern flank of this mountain, there is an extensive sequence of moraine ridges which indicate the former presence of a small local ice mass. However, there is no cirque form in which a glacier could develop and the backwall is virtually straight with cliffs which are 100m in height for a horizontal distance of 1km. This cirque glacier clearly demonstrates the relationship between landslides and local glaciation (Fig. 30).

Site Description

A series of moraine ridges can be identified about 2-300m from the backwall. The lowest altitude of the moraine is 530m OD and orientation of the former ice mass, taken at right angles to the strike of the backwall is 57°. The moraines are furthest from the backwall in the central part, around Windy Hills (SD 76159675), where there is only one ridge, breached by Low Soursike which probably acted as a proglacial meltwater channel when the ice mass extended to the moraine ridge. Further south, there are a number of fragments of moraine ridges which eventually terminate without curving back towards the cliff at Far Cote Gill (SD 76409626). The distal slope of these moraine ridges south of Low Soursike towards Far Cote Gill is marked by a number of discrete slope failures. North of Windy Hills, the ridges are more numerous, but turn sharply towards the cliff base and are smaller and form a more complex pattern which terminates at a small lateral moraine ridge at 600m OD (SD 75679695). There are three main ridges which are separated by enclosed hollows. Mapping has clearly shown that in this area, these moraine ridges lie on top of a large slope failure which extends 200m downslope from the outer moraine ridge (Fig. 30).

Slope failures have also been identified outside the lateral moraine which defines the northern boundary of the former ice mass. Upslope of the lateral moraines, there

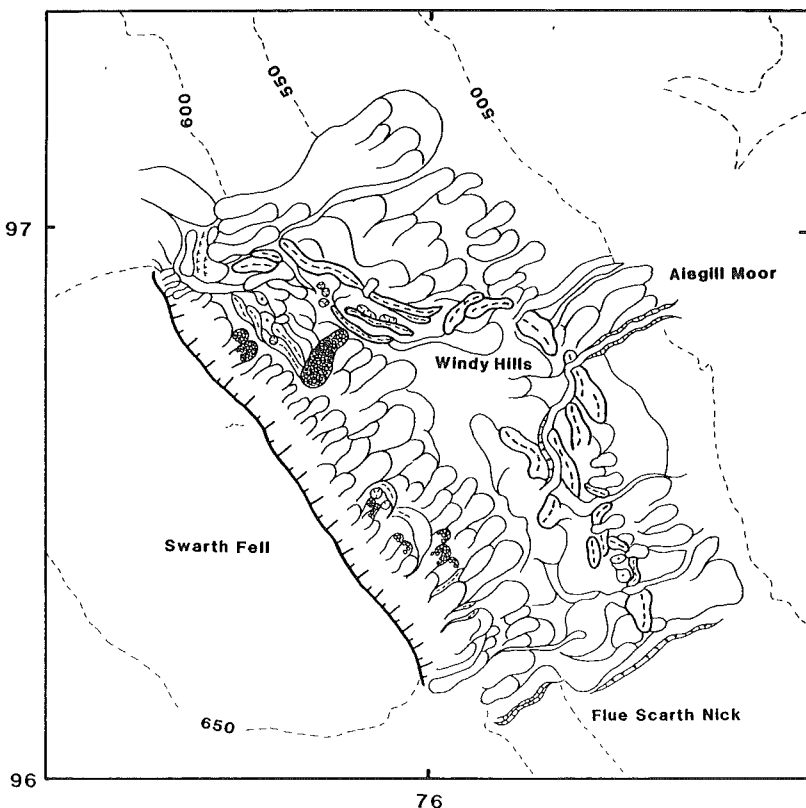


Figure 30. Geomorphological Map of Swarth Fell. Scale and orientation given by National Grid coordinates (from Mitchell, 1991).

is a large rotated block with an associated graben structure indicating rotational, deep seated failure of this part of the mountain (Fig. 30). Within the moraine limit there is a further complex sequence of mass movements along the back wall. At the northern end of the cliff, there has been a deep seated failure which has resulted in the rotational and subsequent flow of the material as indicated by the distribution of grit boulders in a series of flow lobes at the base of the cliff. The whole of the cliff is marked by numerous lobate forms, some of which are composed of grit boulders (Fig. 30) indicating failure and flowage.

Towards the southern end of the cliff there are two small asymmetric ridges which enclose small hollows developed on the lower part of the cliff, but below the free face (SD 759965). Their form and location would suggest that they are

protalus ramparts, although they may also be blocks of bedrock which have slid down the backwall with minimal rotation.

The geomorphic evidence clearly shows that there are a number of slope failures which predate the reoccupation of the 'cirque' by a local glacier, that further mass movements occurred in juxtaposition with the former glacier and that since the disappearance of the glacier, slope instability has continued within the former glacier limit. Although there are no dates available for Swarth Fell, from the limited dates which are available, this suggests a sequence of slope failure activity from the Lateglacial Period, through the Loch Lomond Stadial when the glacier came into existence, and through into the Holocene.

Palaeoclimate

The calculated mean ELA value is 567m (Table 5). Snowblow factor is not seen as an important control on this glacier (Fig. 24; Table 6), with the highest value for potential snowblow factor in the northwest quadrant. The existence of this palaeoglacier appears to be associated with high ground which allowed sufficient snowfall without large snowblow areas, with a topographic depression at altitudes which minimised ablation from the glacier. This suggests that the ELA is a regional characteristic rather than caused by local environmental controls.

Glacial Reconstruction

Reconstruction of the Swarth Fell palaeoglacier (Fig. 31) gives an ice mass which is 0.37 sq. km in area, with a maximum ice thickness of 34m (mean 14.6 ± 9.0) and a mean volume of 0.005 km^3 (Table 3). This is a very thin glacier with a broad rather than long plan outline. Calculations of basal shear stress and ice surface velocity for this former ice mass give the lowest values for the five glaciers identified in the area (Fig. 25). Values for ice velocity (Table 7) indicate that the glacier was scarcely moving, yet it managed to produce a series of terminal moraine ridges. In an attempt to resolve this problem, the former glaciers have been reconstructed for an 'advance mode' (cf Chapter 4). By increasing the curvature of the former ice surface (Fig. 32) this thickens the glaciers and therefore increases values of basal shear stress and ice velocity. However, although this results in a large percentage increase in glacier velocity, it is still a very slow moving ice mass

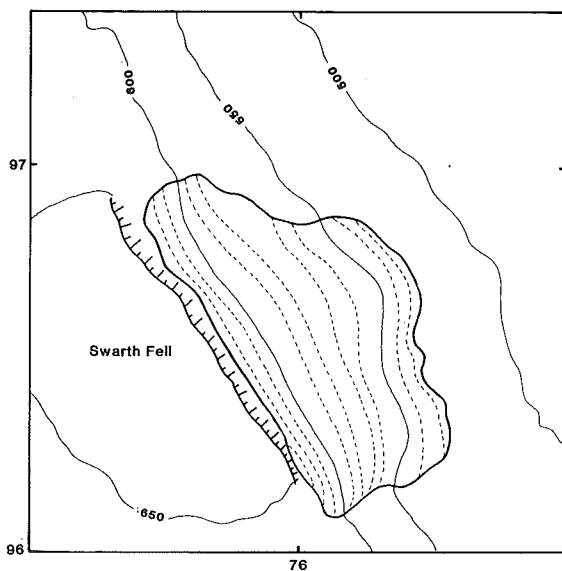


Figure 31. Reconstruction of the Swarth Fell palaeoglacier for 'normal' mode (in mass balance equilibrium); former surface contoured at 10m intervals (from Mitchell, 1991).

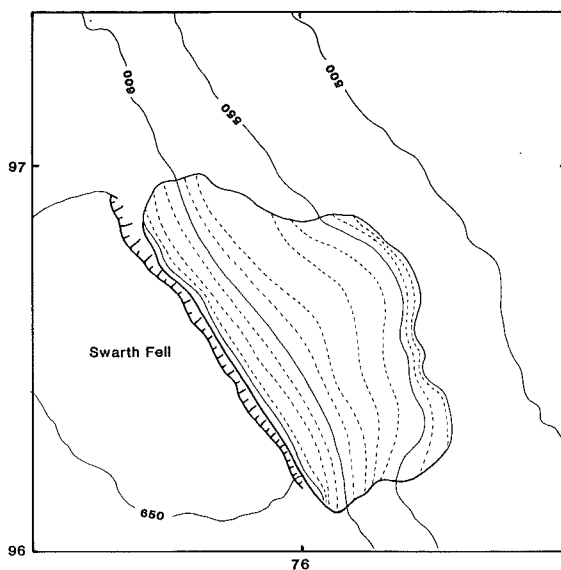


Figure 32. Reconstruction of the Swarth Fell palaeoglacier for 'advance' mode (thicker glacier towards snout); former surface contoured at 10m intervals (from Mitchell, 1991).

(Table 7). The formation of the large terminal moraine ridges remains a problem which may be partly explained by the low shear strength of the materials, mainly mudstones, which were able to be deformed and moved by even this small ice mass. It also seems probable that the back wall would have experienced landslides prior to the development of the glacier as demonstrated at the northern end, which could have provided debris for subsequent glacial transportation.

WILD BOAR FELL

W.A. Mitchell

Introduction

Wild Boar Fell (708m) (SD 758988) lies on the western side of Mallerstang, the upper valley of the River Eden. On the eastern flank of the mountain there is a complex topography of numerous ridges, mounds and boulder-strewn terrain which extends from the valley of Ais Gill (SD 765972) in the south to the low ground east of White Walls in the north (SD 764995).

Wild Boar Fell forms the northern part of an outlier of Namurian strata and is composed of two major rock types - mudstone and coarse sandstone (grit), with secondary lithofacies of cherts and ganister (Dakyn, *et al.*, 1891; Rowell and Scanlon, 1957a; Ramsbottom, 1974b). The lower part of the sequence is dominated by thick mudstones with occasional thin cherts and thin limestones which are overlain, towards the top of the mountain, by a number of distinct sandstone or ganister beds. This stratigraphic situation is particularly significant with respect to slope stability, since it places rocks of high rock mass strength above low strength lithofacies which will fail given suitable geotechnical conditions (*cf* Selby, 1987). This instability is further enhanced by the presence of discontinuities within the rock mass in terms of prominent joints and the presence of a number of small faults have also been mapped towards the northern part of the mountain (Dakyn, *et al.*, 1891). To the north of Wild Boar Fell, a major fault, the Stockdale Vein (or High Dolphinsty Fault), crosses Mallerstang downthrows the Namurian strata by c. 300m on its northern side towards Little Fell (Dakyn, *et al.*, 1891).

Detailed mapping of these complex and spatially confined landforms has been completed using acetate film on enlargements of aerial photographs at an approximate scale of 1:3,000 and research is still at a preliminary stage (Fig. 33; Fig. 34) and no work has yet been done on the geotechnical properties and ages of these landslides. The area will be described from south to north, with the main cliff of Wild Boar Fell divided for ease of discussion with respect to the different names given to parts of the main escarpment called scars.



Figure 33. Detailed geomorphological map of the mass movement forms below Low White Scar and High White Scar, south part of Wild Boar Fell; mapped at an approximate scale of 1:3,000 using enlargements of aerial photographs.

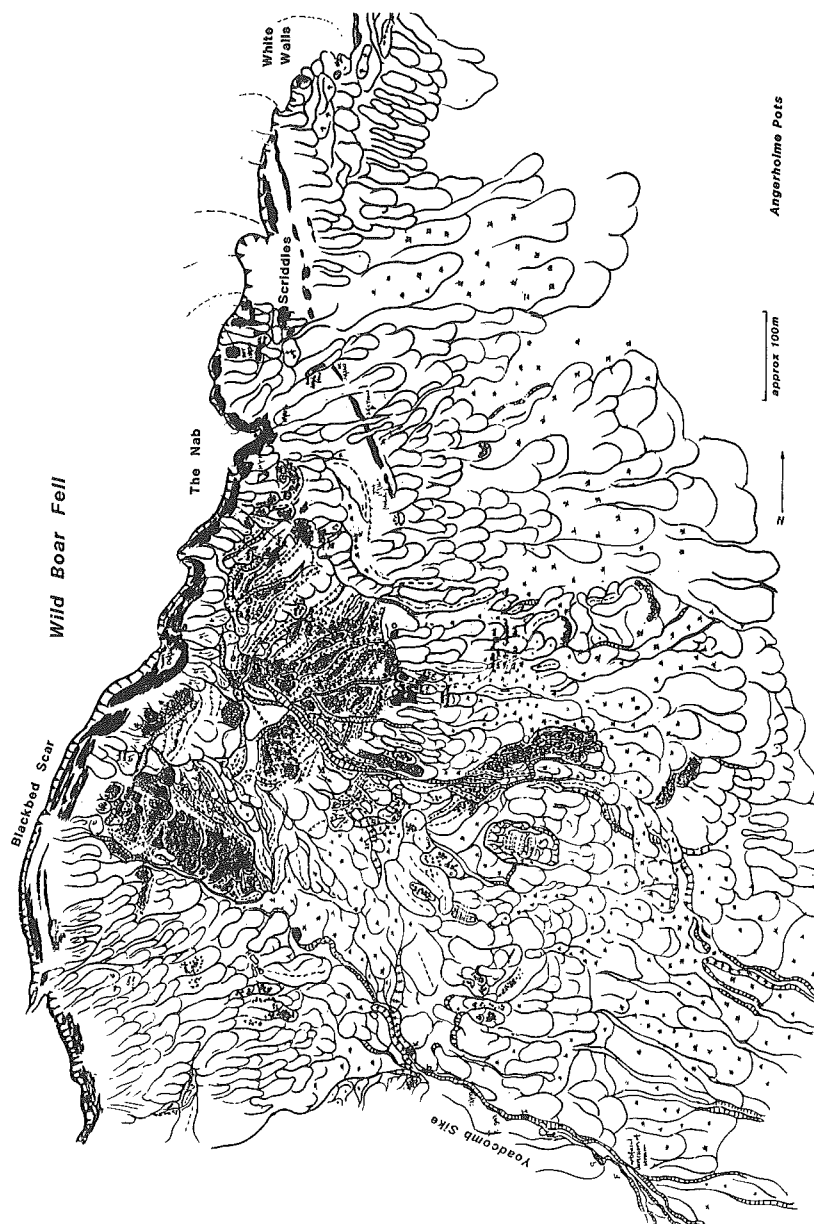


Figure 34. Detailed geomorphological map of the mass movement forms below Yoadcomb Scar, Blackbed Scar and Scriddles, north part of Wild Boar Fell; mapped at an approximate scale of 1:3,000 using enlargements of aerial photographs.

Low White Scar (SD 761978)

This edge forms the most southerly of the escarpments and is at a lower altitude (600m) to the main plateau edge (700m), which is significant because it means that the lithological sequence is mainly mudstones, with only the lowest gritstone exposed in the cliff face. This dominance of the mudstones is reflected in the style of mass movements which have occurred below Low White Scar southwards towards the valley of Ais Gill (Fig. 33). This part of the mountain is characterised by numerous small ridges/terraces which are the upper surface expression of shear planes within the thick shales which occur along this slope. Small graben structures and reversed slopes within this area indicate that extension and rotational slippage have occurred but this is on a small scale because of the high density of discontinuities and resultant failure planes within the weak strength rock. This area of incipient failure extends over 500m from the cliff face and extends as far as Ais Gill. The lower part of the rocks shows a high amount of deformation (downslope creep) and a number of discrete rotational failures with toe lobes which have moved, probably by flow, into the Ais Gill valley since the debris can be seen to lie on top of older failures in the channel side (Fig. 33).

High White Scar (SD 761983)

This escarpment is separated from Low White Scar by an area of tension gashes which indicates the lateral margin of larger mass movements which have allowed High White Scar to retreat about 100m further west than Low White Scar (Fig. 33). The slope failures within this area are much more complex, with a sediment input of gritstone blocks from a higher lithostratigraphical position of the backwall. Furthermore, the mass movements are stacked over a horizontal distance of 800m and vertically over a range from 450m to 700m. The older, and more extensive failures have occurred within the mudstones, and resulted in large deep seated rotational mass movements which have depositional zones which are characterised by numerous secondary smaller rotational failures and debris flow lobes (Fig. 33). On the top of these slope failures are a number of landforms which are defined by large impressive lateral ridges, or levees, and which are composed of large angular blocks of sandstone (gritstone) (Fig. 33). It is these ridges which were interpreted by Rowell and Turner (1952) as lateral moraine ridges. There are a number of ridges, 2-4m in height, of which the most impressive extend 3-400m directly from the present cliff downslope, and define a number of discrete failures which have flowed

downslope. These have occurred at different times since they have been overrun by subsequent flows as indicated by the presence of boulders of younger flows on the levees. However, whether this flow was as a rock glacier or a large debris flow is difficult to ascertain and this research is currently being developed, since their origin has implications for both their environment of formation and their age.

Yoadcomb Scar (SD 761986)

Because of faults which alter the position of the sandstones, this part of the cliff face shows evidence of numerous flow lobes but with few boulders. The scar forms the upper 20m of the cliff with a further 80m of steeply inclined mudstones to the base at Yoadcomb Hole (SD 763986). It is the first part of the escarpment edge to show a change of direction which defines an incipient 'cirque' form (Fig. 34). This actually marks the crown scarp and bowl of a very large deep seated rotational failure with a crown scarp formed by Yoadcomb Scar and Blackbed Scar and with the main mass of displaced bedrock being moved from Yoadcomb Hole (SD 763986) and Black Bed (SD 764988). The initial failure, which is difficult to delimit because of subsequent failures, has a lateral extent of 900m, a width of 3-400m and an altitude range of 250m and occupies the ground immediately to the north of the stream of Yoadcomb Sike (SD 765985). It is overlain by smaller discrete rotational failures, debris flows and toppling of some of the displaced rock masses and has been overridden in to the north by a further large deep seated rotational failure. The landforms within the northern part of the bowl under Blackbed Scar are discussed with respect to that scar (Fig. 34).

Blackbed Scar (SD 762989)

This is the continuation of Yoadcomb Scar and is defined to include the slope failures which lie to the north of the upper reaches of Yoadcomb Sike, but within 250m of the backwall. Just north of the stream, there is a well defined spread of sandstone boulders which have originated from the southern part of Blackbed Scar and moved downslope over 70m of altitudinal range. The boulders extend for c. 300m north of Yoadcomb Sike and have a width of 80m and are clearly contained within a well defined compression ridge/lateral levee system which can be mapped from the lowest extent of the boulders along its northern margin. (Fig. 34). Within the boulders a number of boulder lobes and tension gashes can also be observed; these suggest that movement has occurred at different times. The origin

of this feature is open to discussion.

The Nab (SD 764992)

The 400m of the escarpment north of Blackbed Scar and southwards from The Nab is not named but will be described under this heading. The area below the scar is known as Yoadcomb Hill (SD 766988) and is dominated by a complex extensive deep seated rotational failure which is defined by a lateral shear zone which extends from the northern part of Black Bed eastwards for 800m downslope. This failure is also marked by conspicuous levees, over 400m in length, on both sides of the feature extending from the degradation zone to the accumulation zone (Fig. 34). The lower part of this failure, is marked by a well defined lobate area of sediment which extends down to 470m OD. The upper part of the failure shows a number of tension gashes and graben structures with rotational displacement of bedrock and with a general covering of sandstone boulders. Multiple secondary failures characterise the lower depositional zone and track of this large mass movement. This is quite clearly a large rock slope failure with deep seated rotation along a number of shear planes, although depth to shear plane has yet to be established. It is overlain by a number of smaller failures and rockfalls with ridges which could be interpreted as protalus ramparts at the base of the crown scarp (Fig. 34).

Scriddles (SD 764993)

This is the area of the escarpment to the north of The Nab where a fault with a downthrow on the northwest side (Dakyn, *et al.*, 1891) means that this part of the backwall is mainly in mudstones. This is reflected in the large debris flow lobes which occupy the back wall and the low ground to the east, extending over 500m downslope to an area of limestone pavement (Angerholme Pots) which is caused by an outcrop of the Main Limestone (Fig. 34).

Discussion

All of the features described above were used by Rowell and Turner (1952) to define the extent of a small cirque glacier on the eastern flank of Wild Boar Fell. It has been demonstrated that all of the features can be more easily explained as due to a variety of slope failure mechanisms (cf Chapter 4). The possible explanation of some of these landforms as of periglacial origin as either rock glaciers or protalus ramparts

is an important factor which is currently being investigated. There are no dates on these particular features, but a pollen core from a small lake impounded by a mass of bedrock formed by rotational failure at Coalwell Scar (NY 796014) on the other side of Mallerstang only recorded a Postglacial sequence (postscript, in Rowell and Turner, 1952; p. 208). Since this is close to the present escarpment and in a similar topographic position to the upper slope failures described for Wild Boar Fell, it suggests that some of the landslides may be of Loch Lomond Stadial age. However, the more extensive features are probably associated with paraglacial readjustment during the latter part of the Dimlington Stadial, as indicated by the landslide debris under a terminal moraine ridge at Swarth Fell (Rose and Mitchell, 1989; Mitchell, 1991; this guide). If the features near the backwall reflect periglacial conditions during the Loch Lomond Stadial, then it suggests that the main failures are of Lateglacial age and that there has only been restricted slope failures, mainly rockfall, during the Holocene.

The antiquity of many of the landslides in the southern part of the Pennines has been demonstrated by a number of studies which suggest that mass movement activity occurred over long periods of time (for example, Johnson, 1987; Johnson and Vaughan, 1989; Skempton, *et al.*, 1989). Interdigitation of periglacial head deposits with landslide debris (Johnson and Vaughan, 1989) suggests that the slopes were unstable under periglacial conditions which would have reduced rock mass strength and initiated failure. The next part of the investigation of these mass movements will be directed towards finding sites where soils have been overridden by the failed material and attempt to establish a chronology for the mass movement events.

GRISEDAL

W.A. Mitchell

Introduction

Grisedale (SD 7793) lies between the northern and northeastern flanks of Baugh Fell (SD 7291) and Swarth Fell (SD 7596) (Fig. 1). The dale is drained by Grisedale Beck which rises on East Baugh Fell and becomes the River Clough when it enters Garsdale at Clough Dub (SD 783922). Grisedale is a critical area in the determination of the two former ice flow events, since there are a number of drumlins and superimposed drumlins within the dale.

Superimposed Drumlins

A wide range of drumlin forms have been mapped within Grisedale (Fig. 35) as far west as Holmes Moss Hill (SD 753946). They range from a sequence of four thin parallel drumlins south of Scale (SD 764938) to more massive features further down the dale (SD 768936). The drumlins are continuous eastwards with drumlins on Garsdale Low Moor (SD 789928) and southwards with drumlins in the upper part of Garsdale as far west as Raygill (SD 765901) (Fig. 35). Summary statistics for both the 60 drumlins in Grisedale and the 91 drumlins in Garsdale are given in Table 10. The overall pattern of former ice flow indicated by these drumlins is of convergence of ice from Grisedale, East Baugh Fell and Garsdale in the vicinity of Moorcock Inn (SD 799927). This indicates that ice flowed up Garsdale and was contiguous with ice flow eastwards into Wensleydale reflecting a much more extensive ice centre than previously envisaged.

		ORIENTATION	LENGTH	WIDTH	AREA	ER	K	Sd
		(°)	(m)	(m)	($\times 10^{-4}$ sq km)	dimensionless	dimensionless	dimensionless
GARSDALE (91)	\bar{x}	82.6	321.1	136.4	350.6	2.38	2.57	0.43
	range	35-124	125-600	70-270	67-1051	1.14-3.63	1.31-5.65	0.16-0.86
GRISEDAL (60)	\bar{x}	114.9	267.4	134.9	287.6	2.11	2.28	0.40
	range	42-171	110-580	65-320	53-1142	1.22-4.22	1.29-4.77	0.11-0.72

Table 10. Summary statistics of drumlin morphology in Grisedale and Garsdale.

Superimposed drumlins have been mapped on the lower northeast flank of Baugh Fell in the recently forested area between Mouse Syke (SD 772928) and Crookshaw Gill (SD 767935) (Fig. 36). Mapping of these particular drumlins indicates that the base lines of the megadrumlins and superimposed drumlins are often discontinuous, particularly with respect to the breaks of slope of the smaller forms. Whilst the overall profile of the megadrumlin may be apparent, identification of the crest line of the megadrumlin may be difficult having been destroyed by later flow events. The high point of the megadrumlin is usually shared with the high point of one of the superimposed forms.

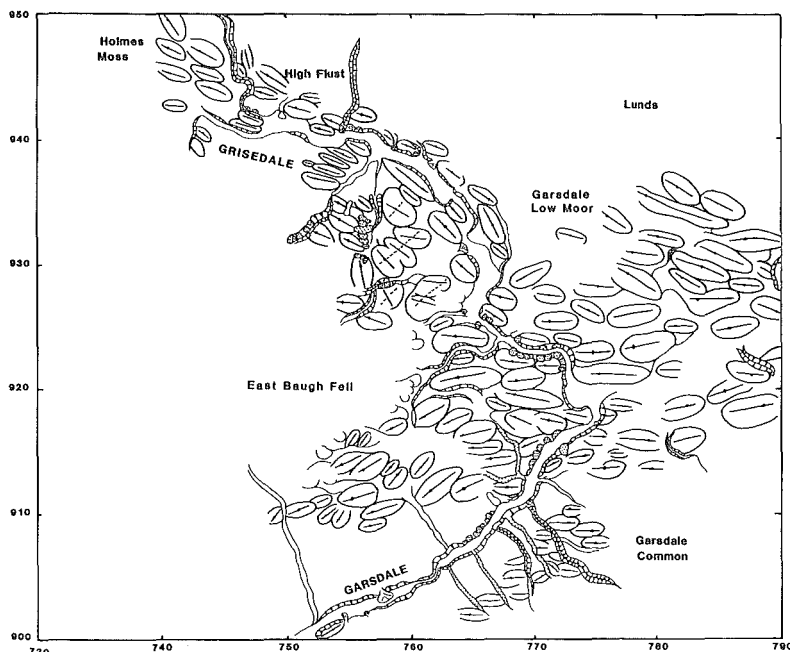


Figure 35. Geomorphological map of the drumlins in Grisedale and Garsdale Head (from Mitchell, 1991). Crest line of original drumlin given by dashed line. Scale and orientation given by National Grid coordinates.

Superimposed forms are most clearly observed where there has been a distinct change of bedform trend between the flow direction inferred from the trend of the larger drumlins or megadrumlins, and the later ice flow which created the superimposed forms. This is the situation in Grisedale, where mapping of the drumlins around Butterbeck (SD 7793) revealed that certain large drumlins could

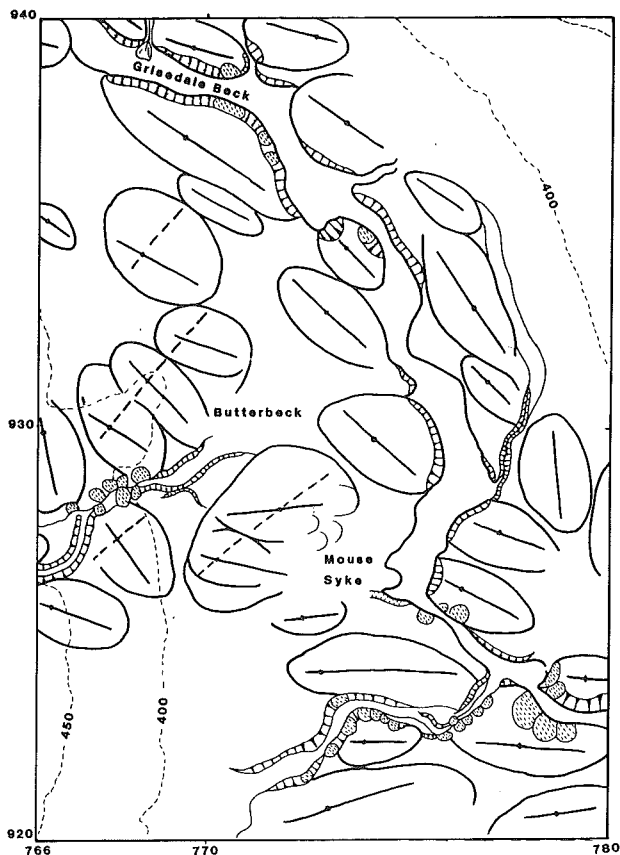


Figure 36. Geomorphological map of part of Grisedale showing the superimposed drumlins around Butterbeck. Crest line of original drumlin form given by dashed line. Scale and orientation given by National Grid coordinates.

be identified which were orientated across the valley. Superimposed upon these drumlins are smaller forms which are orientated at an angle to the first landforms and indicate a later flow in a down valley direction (Fig. 36).

These different drumlin forms indicate two flow events (cf Fig. 12 and 13); during the first (earlier) event ice flow was off the eastern side of Baugh Fell across Grisedale towards Aisgill. This was replaced at a later stage, by ice flowing down Grisedale, from a source area across Baugh Fell to Swarth Fell. Whilst this later flow modified the drumlins in Grisedale, it did not alter the drumlins which occupy the ground between Grisedale and Garsdale where only eastward trending drumlins are found on Garsdale Low Moor (SD 785927) (Fig. 36).

LOVEN SCARS AND HANGINGSTONE SCAR

W.A. Mitchell

Introduction

The major escarpment on the eastern side of Mallerstang is divided into a number of distinct topographic units. This part of the guide deals with the southern part of the escarpment which includes High Loven Scar (NY 799001), Low Loven Scar (NY 797001) and Hangingstone Scar (SD 798995), which have been mapped in a similar manner to Wild Boar Fell. These show a complex pattern of spectacular mass movement phenomena, particularly mudslides (Allison and Brunsden, 1990), which are spatially interlinked. The northern part of the escarpment, Mallerstang Edge (NY 798015) shows similar landforms and will be mapped in the near future. These features have also been interpreted as moraine ridges in the literature (Rowell and Turner, 1952) but can again be demonstrated to reflect multiple slope failures, particularly mudslides and rotational failures. As with the mass movements on Wild Boar Fell this research is at a preliminary stage and no discussion is made regarding the geomechanical conditions which may have been responsible for the generation of these different slope failures. Their possible age has been discussed with respect to Wild Boar Fell.

These scars reflect the lithological control of massive sandstones overlying a thick shale sequence. There is greater structural displacement along this escarpment than on Wild Boar Fell, with a high number of west-east and northwest-southeast faults (Rowell and Scanlon, 1957a). The main displacement is the Stockdale Vein which downthrows the strata by over 100m and extends east-west across the northern part of the Askrigg Block. In Mallerstang, it occurs as two major faults on either side of Hugh Seat (Dakyns, et al., 1891; Rowell and Scanlon, 1957a). The southern fault is also known as High Dolphinsty Fault and can be seen crossing the valley just north of Wild Boar Fell. The main displacement occurs just north of High Loven Scar where there is a major indentation in the escarpment since the discontinuity has been exploited by slope failures (cf Fig. 2).

The lithology can be generalised as a coarsening upwards sequence which forms a sequence of five distinct lithofacies in the main cliff face of High Loven Scar and

Hangingsone Scar, which is part of the lithostratigraphy known as the Fossil Sandstone (Dakyn, et al., 1891; Brenner and Martinsen, 1990) which lies stratigraphically above the Namurian rocks of Wild Boar Fell. The main rock face is composed of a number of different sandstones interstratified with shales (Table 11). The sandstones often have a channel architecture and are often not continuous along the cliff sections. A generally thick shale sequence underlies the sandstones but is not widely exposed (Dakyn, et al., 1891).

The scalloped nature of the main escarpment indicates that there have been numerous failures along its length, creating a number of different crown scarps associated with curved shear planes. These deep seated failures have produced a major sediment input to slope instability which has been transformed into a complex sequence of mudslides downslope.

Facies	Lithology	Structures and Fossils	Degree of Bioturbation
Shale	Clay-rich shale with thin beds of siltstone and v.f.-f.-gr. sandstone	Mottled, wavy lamination	Poorly exposed
Sandstone Shale	V.f.-f.-gr. sandstone with thin interbeds of shale	Ripple laminated - mottled, discontinuous thin beds	Low to moderate
Wavy-Sandstone	V.f.-f.-gr. mica-rich sandstone with some thin shale laminae	Wavy-ripple lam.; flaser beds, rare burrows	Low
Tabular Cross-Bedded Sandstone	F.-m.-gr. mica-rich cross-bedded sandstone; tabular foresets with angular contacts; conc. of shale chips and fossil fragments along basal erosional contact	Tabular cross sets up to 1.5 m thick; strata dip eastward 15-25 degrees, erosional set contacts nearly horizontal; fossil fragments: crinoid, brachiopod	None
Mottled Sandstone	V.f.-med.-gr. poorly sorted sandstone	Mottled, indistinct strata with some burrows preserved at top of units: <i>Mono-craterion</i> (<i>Skolithos</i>), <i>Olivellites</i> , <i>Zoophycos</i>	High

Table 11. Lithological characteristics of the main facies of the Fossil Sandstone (from Brenner and Martinsen, 1990). Reprinted by permission of the Yorkshire Geological Society.

Low Loven Scar (NY 797001)

Low Loven Scar (560m OD) exposes the lowest gritstones in the sequence. Dip measurements show low values ($< 5^\circ$) suggesting that the rock is in situ, although there are higher values ($> 30^\circ$) along the northern flank, where the deep seated movements which originated along the northern part of High Loven Scar/High Band (NY 79950021) have started to affect this lower escarpment. It can be shown that this scar has generated its own failures, particularly multiple flow events,



Figure 37. Detailed geomorphological map of the mass movement forms below Low Loven Scar, High Loven Scar and Hangingstone Scar; mapped at an approximate scale of 1:3,000 using enlargements of aerial photographs.

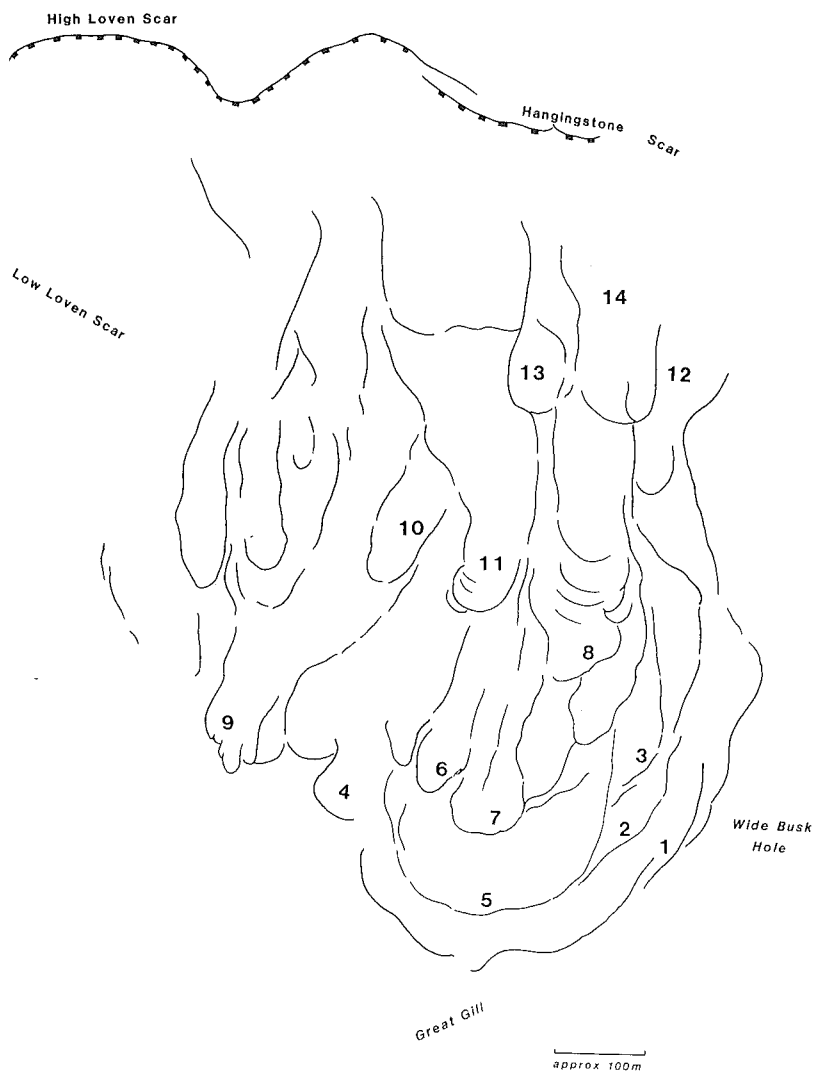


Figure 38. Number of mudslide events below Hangingstone Scar identified from the detailed geomorphological mapping of Figure 33. Events are numbered from oldest (1) to youngest (14).

with the upper flows incorporating large numbers of sandstone blocks (Fig. 37). These extend about 250m downslope, with a breadth of between 50 and 100m. The lower slopes towards Knowes, between 480 and 400m OD, have a high density of smaller debris flows which extend 700m downslope from the cliff, with numerous tension gashes crossing the slope (Fig. 37), similar to the imbricate pattern identified on Low White Scar, on Wild Boar Fell. The area to the north, towards Wether Hill (NY 792006), is also covered by numerous features associated with mass movements, particularly mudslides, from a source area to the north between High Loven Scar and High Band (NY 799003). Slope failures from the western part of High Loven Scar also show signs of influencing Low Loven Scar by an input of debris at either end and over the cliff (Fig. 37); there is also evidence for present activity of material over the cliff.

High Loven Scar (NY 799000)

This marks part of the main escarpment with heights of c. 660m OD. On the OS map, this scar occurs in two parts with a northern part which is part of the source area for the complex mass movements which have occurred to the north of Low Loven Scar (not mapped on Fig. 37). Along the base of this northwest orientated rock face, there are ridges which could be interpreted as protalus ramparts. The second area is orientated west and forms the source area for material which has been transported to the south side of Low Loven Scar (Fig. 37). The upper part of the slope is characterised by large spreads of gritstone boulders which lie on top of rotated blocks of bedrock which are dipping at angles $>50^\circ$ into the hillslope. These are defined on their southern side by a number of ridges which delimit the lateral margins of a complex of debris flows and mudslides which have been mapped from below the cliff extending 800m downslope (Fig. 37). Many of the individual flows are defined by distinct levees with the former flow surface picked out by sandstone boulders. There are also ridges which lie on top of older ones and which reflect overriding of the levees by later events.

The upper part of these failures which occur on the ground between the top of the cliff of Low Loven Scar, and the base of High Loven Scar, is characterised by a complex pattern of tension gashes which indicate a general zone of extension (Fig. 37). It is not known whether these are presently active or reflect features related to past events which are distributed across the hillslope. Debris transport from High Loven Scar is indicated as having taken place over distances $>800\text{m}$ from the present

crown scar, but there is evidence of multiple secondary failures which have broken away from material which had previously failed and moved downslope. These have then overridden the tracks and accumulation areas of earlier slope failures to produce a vertical stacking of debris lobes (Fig. 37).

Hangingstone Scar (SD 798995)

The ground below this particular part of the escarpment is marked by some of the most spectacular mass movements in the area. Extending >800m from the backwall are a series of very large mudslides, which are characterised by the movement of softened clay rich sediments which has moved across a basal shear surface (cf Brunsden, 1984; Hutchinson, 1988; Allison and Brunsden, 1990). At Hangingstone Scar, they are different from the other slope features in that the toe lobes of the accumulation areas of the largest of these landforms are marked by a distinctive geometric series of curved compression ridges which have distal slopes of >30° and with linear patterns of boulders in the intervening ground (Fig. 37). These ridges are caused by compression as the mudslide loses momentum. Longitudinal shears and radial crevasses are other common features of the mudslide accumulation areas (Brunsden, 1984) and can be identified on Figure 37. The main feeder tracks above these compression ridges are marked by well defined levees which extend some distance along each mudslide with the large number of such ridges reflecting the multiplicity of mudslides which are stacked on this part of the hillside (Fig. 38).

At the head of the mudslides, there are many rotated and displaced bedrock blocks which indicate that the major source of debris to the mudslides has been by deep seated rotational failure of the back wall aided by a near vertical joint set which has determined failure pattern. For example, a well developed graben structure and large rotated block, dipping 26° at 136°, occurs under the crown scarp at the southern edge of the present mapping (Fig. 37). The upper parts of the slope are also characterised by numerous fossil tension gashes which reflect the relative movement of different parts of the source area under tension with different mudslide events.

Although most of the ridges in the accumulation zone reflect both rotated and disaggregated mudrock, there are also some constructional forms. For example, there is a distinct curved ridge about 200m from the present backwall. Its character and location suggest that this is possibly a protalus rampart, and that its

position relative to the backwall now reflects subsequent destruction and lateral migration of the escarpment. Alternatively, it is the frontal lobe of a debris flow. However, it is truncated on its southern side by more recent mass movements which have destroyed the ridge, and therefore gives some indication of temporality in that it is younger than the larger mass movements on which it occurs, but is older than the mudslides which have been fed by material in juxtaposition to the ridge and which have destroyed the southern part of it.

A number of the mudslide tracks also show numerous tension gashes; these are interpreted as Reidel shears associated with differential motion between the plug flow of the main movement and the lateral drag which is reflected in the levees on either side of the flows (Brunsdon, 1984; Johnson, 1984). Complex flow events are also recorded by the pattern of compressive ridges where concave downslope patterns reflect reactivation of the accumulation areas by subsequent events which have moved down the northern flank of the main feature (Fig. 38).

Many of the upper parts of the different mudslides below Hangingstone Scar are covered by a surface layer of large sandstone blocks. These can be used to distinguish different flow events since they reflect individual sediment inputs from the source area into the numerous feeder tracks of the mudslides (Fig. 38).

There are a number of exposures which indicate that under the boulder cover, there is a thin diamict layer which has derived its matrix from the argillaceous rocks and which were responsible for the initiation of slope instability. However it is extremely difficult to generalise, since the diamict composition is highly variable depending on which part of the mudslide is exposed. In the lateral ridges, there are exposures of very fine diamict composed mainly of mudstone, whereas in the boulder field next to that particular ridge, the exposure shows a diamict with large angular grit clasts in a sandy matrix. Exposures in levees along the mudslide tracks show that they are composed of open framework grit boulders with no matrix but can also be composed of diamict with clasts of sandstone and shale in a very compact matrix. There are few exposures in the accumulation areas; one in a ridge near Great Gill (Fig. 37) records clast supported boulder gravel with some sandy matrix, which is adjacent to another exposure which shows a clay rich diamict with mudstone clasts.

The outer lower limit of the accumulation zone extends into the valley of Great Gill (Fig. 37). This stream network may be interpreted as a meltwater system

associated with the last ice sheet, although it seems more likely that these channels were formed by dewatering of the mudslides, which would also have helped them to increase shear strength, lose fluidity and cease movement.

Mapping has been able to distinguish a number of different mudslide events (Fig. 38). This map shows that the different mudslides reflect different source areas along the escarpment and that each of these mudslides appears to get smaller since they do not extend as far downslope as the earlier events.

Discussion

One of the most obvious differences between the two sides of Mallerstang is the difference in styles and types of slope failure. Although there are certain common features, such as the large scale deep seated rotational failure, there are more important differences in terms of the overall geometry of many of the movements. This is clearly seen if the features below Hangingstone Scar are compared to the northern part of Wild Boar Fell. However, it is very difficult to give a detailed explanation of this observation at present, since detailed work has yet to be completed on the sediments which may give some indication of critical differences in lithology and environment, for example in terms of rock mass strength (Selby, 1987) which will have influenced slope behaviour. Furthermore, no detailed morphometric analysis, in the style of Crozier (1973), has been completed for these landslides; as pointed out by Johnson and Vaughan (1989), such an approach may give valuable clues to the former processes responsible for the slope failures.

Studies of landslides in northern England have concentrated on the southern Pennines (cf Johnson, 1987), where they have been treated as individual occurrences and there has been no overall attempt to find underlying regional patterns. However they are also common landforms in this part of the Western Pennines (Mitchell, 1991; this guide). This has not been helped by the fact that many parts of northern England have never been examined for slope failures except in the most general terms or have been misinterpreted as features of local glaciation in Mallerstang (Rowell and Turner, 1952) and on Cross Fell (Johnson and Dunham, 1963). Although the presence of landslides in the latter region has been noted (Tufnell, 1985), this is an area of the Pennines for which little is known regarding its Quaternary history. There is obviously scope for much future research on this topic, both in terms of the mechanics of slope failures but also in

terms of the age of these features and their significance in Quaternary landscape evolution.

BLUECASTER

W.A. Mitchell & T.P. Buggie

Bluecaster is a prominent hill, 345m high, within the Rawthey valley which is composed of resistant dolerite (or diorite in the earlier literature) which has been intruded into the Stockdale shales in the vicinity of the Dent Fault (Underhill, *et al.*, 1988). There are very few outcrops and it appears to be a very small intrusion. In an area such as the Western Pennines, where the majority of rocks are sedimentary rocks, this makes the Bluecaster dolerite a potential rock type for glacial indicator tracing (Saarnisto, 1990), since these erratics can be easily distinguished by mineralogical composition of the rock which gives a weathering corona with a distinctive brown colour.

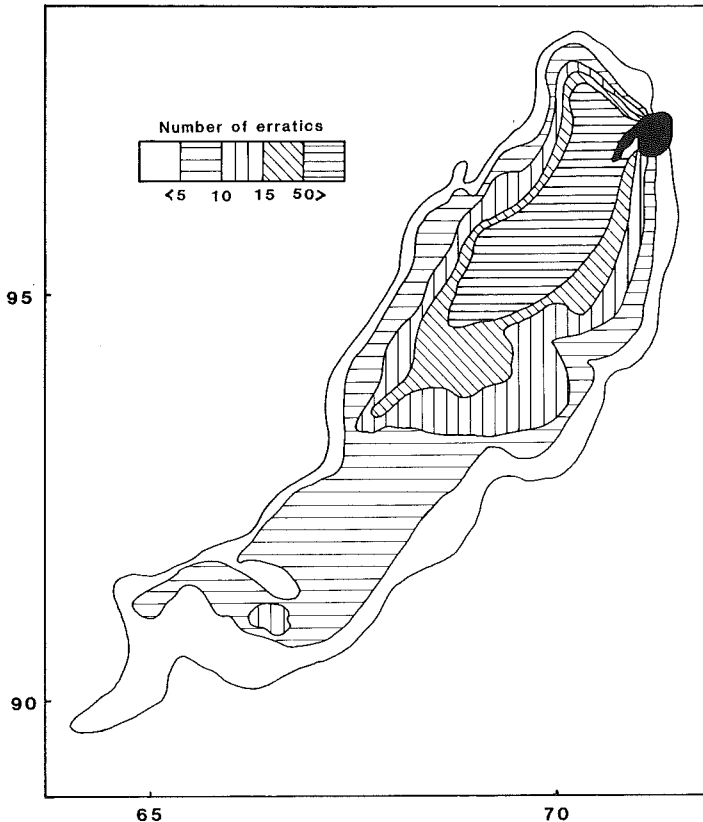


Figure 39. E_N values for the Bluecaster dolerite erratic train. (from T.P. Buggie unpublished). Scale and orientation given by National Grid coordinates.

One method which has been successfully employed in the delimiting of erratic trains is by using the distribution of erratics in random sections of stone walls (Shakesby, 1977; Letzer, 1978). This is based on the assumption that the walls were made of locally available boulders lying on the ground surface so that they have not been removed a great distance from source. The technique requires random samples of wall to be selected and the erratics found in the sample section measured to determine the amount of area of the wall which is composed of the erratic (E_A) and the number of erratics per wall section (E_N) (cf Shakesby, 1977). This information can then be plotted to give an indication of erratic intensity density away from outcrop; this has been done for number of erratics (E_N) (Fig. 39).

The resultant map shows that the main erratic train extends southwest from the outcrop with most erratics occurring within the lowland area of the Rawthey valley. No erratics were found in stone walls to the northeast of outcrop but there is a small

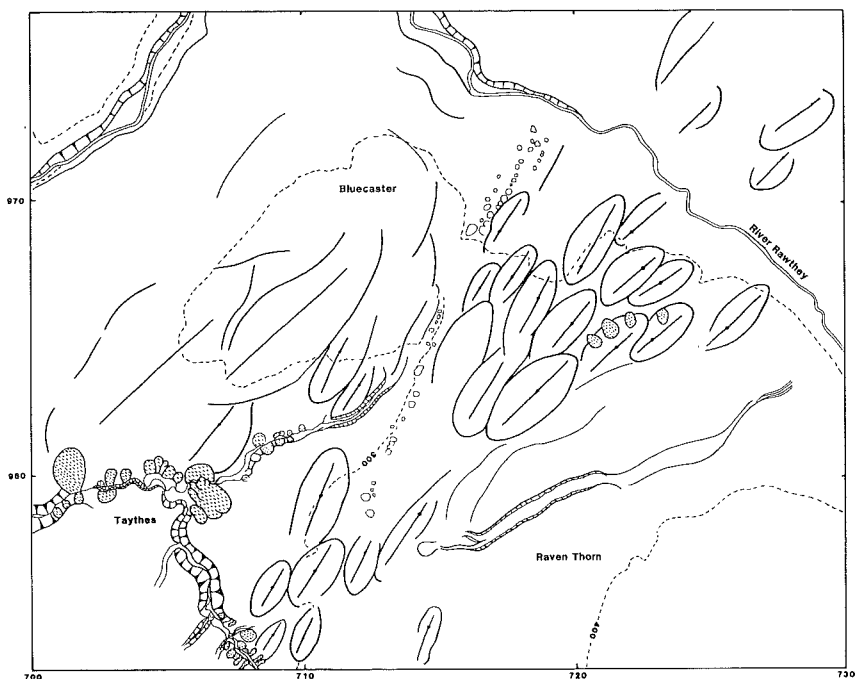


Figure 40. Geomorphological map of the Bluecaster drift tails and drumlins (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

movement of erratics towards the northwest. The erratic train axis is relatively relatively straight and extends for about 10 km and maintains a relatively constant width through its length (Fig. 39).

This information on former ice flow direction is confirmed by the geomorphological mapping of the drumlins and drift tail which extends to the southwest of the diorite outcrop (Fig. 40). This indicates that the ice divide or source area lay to the northeast of Bluecaster on the western flanks of Wild Boar Fell and across the valley of Sally Beck (SD 718985) westwards towards the Howgill Fells. This is confirmed by Letzer (1978) who mapped a series of meltwater channels at Stennerskeugh Clouds (NY 745005) which have an northerly orientation. The more restricted northwest movement of erratics may reflect an earlier flow event where the ice source area lay to the southeast over Baugh Fell, although the presence of dolerite in till sections near Kirkby Stephen (Letzer, 1978) suggests that northwards travel of erratics has also occurred; this is interpreted as occurring within Ice flow event 1 (cf Fig. 12).

CAUTLEY CRAGS

W.A.Mitchell

Introduction

On the northeastern slopes of Great Dummocks (660m OD) a small ice mass has eroded a cirque at Cautley Crag (SD 6896), the only well developed landform associated with local glacial erosion in the eastern Howgill Fells (Fig. 41). This cirque which is cut in Silurian rocks, has a semi-circular plan and is best developed on the western and southern sides where there are cliffs 150m in height. The upper altitude of the cliffs is c. 650m OD and the floor of the cirque is c. 350m before it steeply descends to the floor of Cautley Holme Beck at 220m OD. Orientation of the long axis of the cirque is 29°. Associated with this erosional landform are a number of moraine ridges within the cirque. There is also a large depositional ridge in the valley bottom, to the east of Cautley Spout (SD 683976), below the scree slopes south of Yarlside (Fig. 41).

No detailed mapping of these forms has been previously attempted and earlier studies have all inferred that the large mass of drift in the valley bottom is the terminal moraine ridge associated with a small local ice mass within the cirque (Gunson, 1966; King, 1976; Harvey, 1985). Detailed mapping has shown that the geomorphology of this area is more complex, with the former glacier of more restricted extent and that the drift on the valley floor is a talus foot rock glacier.

Cautley Crag's Glacier

A number of moraine ridges occur at the cirque lip at 340m OD (Fig. 41) and identify the maximum extent of the former glacier and pattern of ice recession. On the eastern margin there are four lateral moraine ridges which extend from 450m OD to 350m OD. The ridges are c. 200m in length and are continued downslope by a longer (300m) fragment of lateral moraine ridge which extends from the innermost ridge. This ridge extends to the cirque lip before curving round as a terminal moraine defining the maximum extent of the former glacier. However, it does not extend along the cirque lip, probably due to the steepness of the slope which did not allow ridge construction (Fig. 41). The western limit is defined by a well developed lateral moraine ridge to the north of a series of mounds on the west side of the main stream within the cirque (Fig. 41).

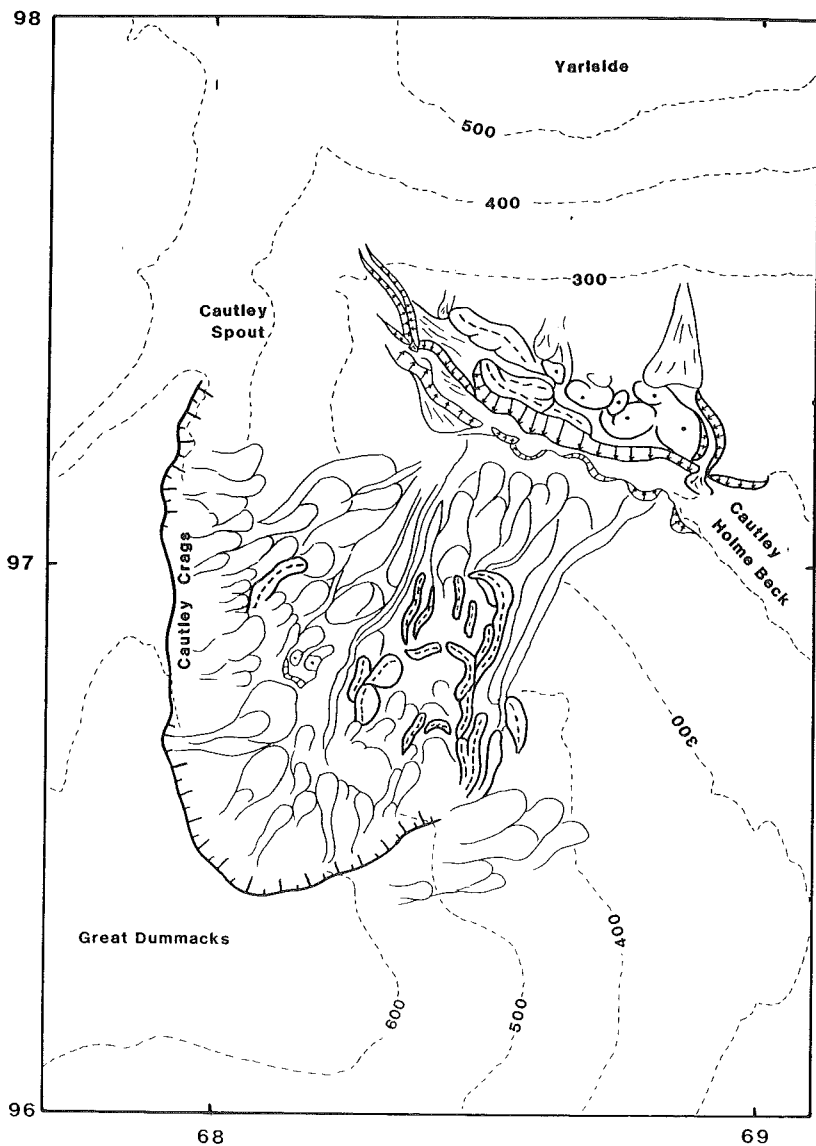


Figure 41. Geomorphological map of moraine ridges below Cautley Crags and the ridge complex in the valley of Cautley Holme Beck (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

Four further fragments of moraine ridges, which are orientated downslope, may also be identified between the eastern lateral moraine and the small stream which drains the cirque and are associated with ice recession. A most complete arcuate moraine ridge can be identified as a lateral margin on the eastern flank of the cirque extending across the cirque floor down to 390m OD, before continuing upslope on the eastern side of the stream as two parallel lateral ridges (Fig. 41). On the western side of the stream, there are a number of small hummocks and a large mound of drift associated with a slope failure. The extent of the former glacier within the cirque is therefore marked by the lowest moraine ridges at the cirque lip and indicates that the ice mass was much smaller than was previously thought (cf King, 1976).

All of these ridges occur in close proximity to a large number of debris flow lobes, talus cones and alluvial fans which extend from below the cliffs or from gullies cut into the backwall of the cirque. Lobes are also common on the steep slope below the moraine ridges and on the slopes of Coonhard Brow to the east of the cirque (Fig. 41). A very large lobe of debris forms a distinct asymmetrical ridge north of the most westerly identified lateral moraine ridge. This has been interpreted as a mass movement feature which occurred at the margin of the former glacier, although it may be a protalus rampart.

Palaeoclimate

The ELA for this glacier gives a value of 424m OD with an altitudinal range from 330m up to 600m, the largest range of the five former glaciers. Total potential snowblow area of 8.1 sq. km lies predominantly in the western sector (226-315°) giving a snowblow factor of 5.03, with a mean snowblow factor of 5.81 (Fig. 24; Table 6). This suggests that westerly winds were responsible for the low ELA of this glacier reflecting the significance of snowblow from the surrounding plateau areas of the Howgill Fells.

Glacier Reconstruction

Reconstruction of this glacier gives a small ice mass of 0.24 sq. km (Fig. 42) with the largest altitudinal range (330 to 600 m) for any of the five glaciers and is also the most northerly orientated of the ice masses. The reconstruction, plus the steeper cirque floor developed on Silurian rocks, gives a much thicker glacier with a steeper ice surface gradient, which gives the highest values of basal shear stress (Fig. 25)

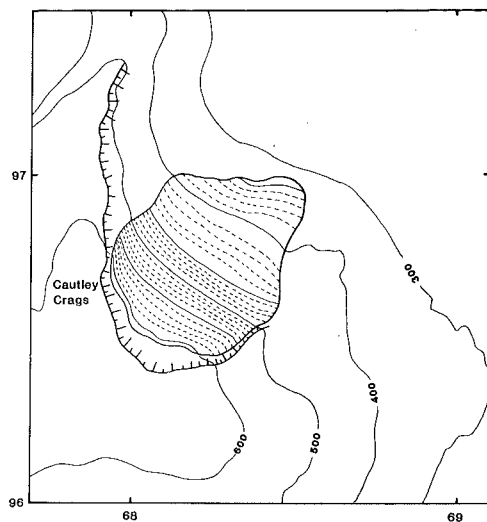


Figure 42. Reconstruction of the Cautley Crags palaeoglacier for 'normal' mode (in mass balance equilibrium); former surface contoured at 10m intervals (from Mitchell, 1991).

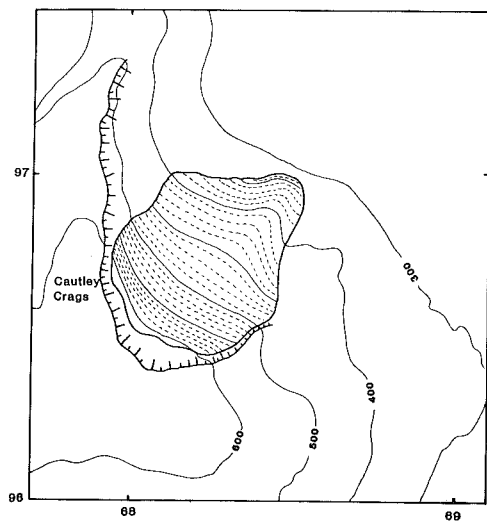


Figure 43. Reconstruction of the Cautley Crags palaeoglacier for 'advance' mode (thicker glacier towards snout); former surface contoured at 10m intervals (from Mitchell, 1991).

but still only moves the glacier at 0.436 ma^{-1} (Table 7). Reconstruction of this glacier in 'advance mode' (Fig. 43) gives a value of 2.01 ma^{-1} which is an acceptable figure at the lower end of a scale for a cirque glacier (Paterson, 1981).

Cautley Craggs Rock Glacier

The mass of drift in the valley bottom extends from the confluence of Cautley Spout Gill and Bowderdale Gill (SD 684875) eastwards for 500m to where an unnamed stream descends off Ben End (SD 690872). The thickness of sediment exceeds 20m and overlies bedrock which is exposed in sections cut by Cautley Holme Beck, on the southern side of the landform (Fig. 39). This feature has previously been interpreted as a terminal moraine ridge deposited by the former glacier which occupied the cirque (King, 1976; Harvey, 1985).

Detailed field mapping has shown that this feature cannot be associated with the cirque to the south, but that it was formed by the accumulation of debris from Yarlside to the north (Fig. 41). The slope above the feature to the north is steep and rectilinear and shows no signs of local glacial erosion. At the western end of this mass of drift, there is an alluvial fan deposited by the two streams which descend to the valley floor. This has infilled the bottom of the valley and is banked against three ridges which are longitudinal to the valley axis. The innermost of these ridges extends for 200m and is arcuate at its western end where it curves into the hillside behind it, where it is overlain by active talus (Fig. 41). To the south, two further ridges, each about 100m in length, are found parallel to the innermost ridge curving into the hillside. The plan form of the ridges, and the fact that they are parallel with the base of the slope, together with the nature of the sediment of which they are composed, suggests that the landform is a lobate or talus foot rock glacier.

These ridges cannot be traced eastwards where they are replaced by a number of large mounds which extend the feature 300m eastwards down valley. Near its eastern limit, the landform is dissected by two unnamed fluvial channels which descend from Yarlside and which have created a large talus cone upstream of the mounds and the whole mass has been modified on its southern margin by Cautley Holme Beck which has created a steep fluvially eroded slope.

GARTHS

A.R. Gunson

Introduction

Garths (SD 635923) and Combe Scar (see below) are two of several sites which were investigated as part of a project attempting to date the nature of ice retreat and environmental change around the Howgill Fells during the Lateglacial Period (Gunson, 1966). Analytical procedures used at each site reflect the techniques appropriate to palynological investigations at this time. For each site, a levelled transect of cores was carried out using a Hiller corer and the deepest one (or that which had the most interesting stratigraphy) was selected for pollen analysis. Samples for the latter were collected at 5 cm intervals, stored in air-tight glass vials and prepared as quickly as possible after collection using sodium hydroxide, hydrofluoric acid, acetolysis, then stained with fuchsin and mounted in glycerine (cf Faegri and Eversen, 1964). Counts were made at 400x magnification and the pollen sum was normally 200 (except where *Corylus* was overrepresented, when it was 150 excluding *Corylus*).

An unconventional local zonation was employed based on lithostratigraphy and numbered downwards from the Lateglacial/Postglacial boundary which is a well defined and documented horizon in pollen stratigraphic studies; the prefix L has been used to indicate a local definition for the zones at Garths Farm (Gunson, 1966). *Betula cf nana* was separated subjectively using the method of Walker (1955a).

Site Introduction

Garths (SD 635923) occupies a long narrow, steep-sided kettlehole 300m by 50m in size situated in pitted outwash gravels of the River Rawthey, close to its confluence with the River Lune about 1km west of Sedbergh (Fig. 7; Fig. 44). It lies at an altitude of c. 105m and is drained by an artificial underground culvert and consists of three separate basins, each separated by low lying gravel ridges, with the deepest part of the basin at 8.6m. Each basin demonstrates a Lateglacial stratigraphy, but in the southern basin (Core H), there is a small central area about 30 m in diameter, with additional lithostratigraphic units (Fig. 44).

A typical sequence consists of the following:

	Postglacial organic sediments
L1	Pasty clay with sands and stones (= Younger Dryas)
L2	Red-brown fine detritus mud (= Allerød)
L3c	Silty clay with some organic content with solifluction gravels at edges
L3b	thin mud lens
L3a	Silty clay with some organic content with solifluction gravels at edges
L4	fine detritus mud with microlaminated mud
L5	Smooth blue clay
	Basal gravel

This shows that the lower part of the sequence, at 540 cm, has an additional organic bed (L3b) of a fissile microlaminated nekron mud which is separated by a layer of inorganic clay (L3a) from the main organic muds of the Allerød Interstadial (L2).

Pollen Analysis

No pollen was obtained from the basal clay (L5), although the sediment is composed of smooth blue clay with no evidence of solifluction gravels in its upper part (Fig. 45). The basal part of the lower organic mud (Zone L4) is dominated by an almost treeless vegetation (tree pollen <10%) with herbs making up more than 50% of the pollen (Fig. 45). The herbs are dominated by *Rumex*, *Galium* and *Thalictrum* which are typical of the Lateglacial Period. Cyperaceae, *Salix* and *Betula cf nana* are all more important than tree birches. Samples from the upper part of this unit indicate changes with a rise in *Juniperus* and *Filipendula* (Fig. 45), signalling the relative decline of Lateglacial herbs and the start of a climatic amelioration. *Betula* then starts to rise slowly to 20% replacing *Juniperus* within the sequence. However, this percentage suggests that trees are only thinly scattered in the landscape.

The middle clay (L3a and c) signals a slight reversion of the vegetation. Tree birches decline to about 14% whilst *Betula cf nana*, *Juniperus* and the Lateglacial herbs all show an increase (Fig. 45). This reversion interval is thought to reflect a climatic deterioration and is also indicated by the solifluction of gravels into the basin edges (Gunson, 1966). In the middle of the clay, there is a thin layer of mud (L3b) and the single pollen count obtained from this sediment shows the highest tree

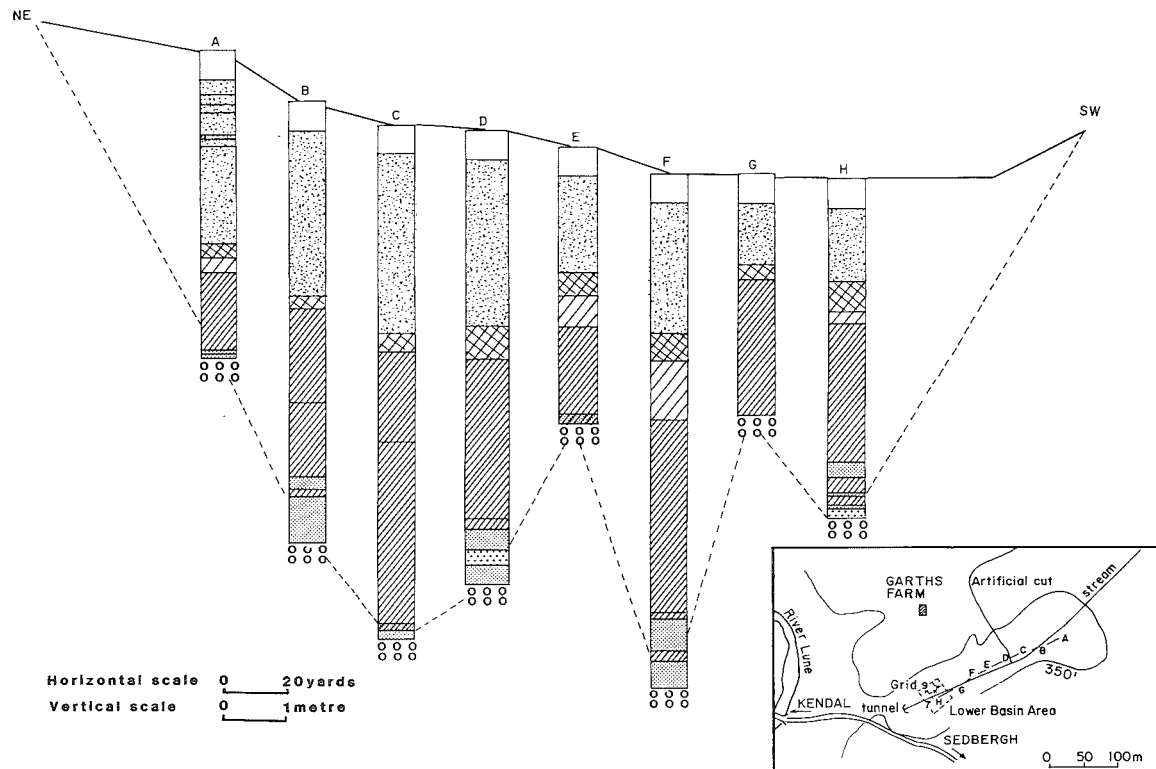


Figure 44. Stratigraphic transect from a series of cores across peat bog at Garths (from Gunson, 1966). Inset shows the location of profile and sites. (For key see Figure 47).

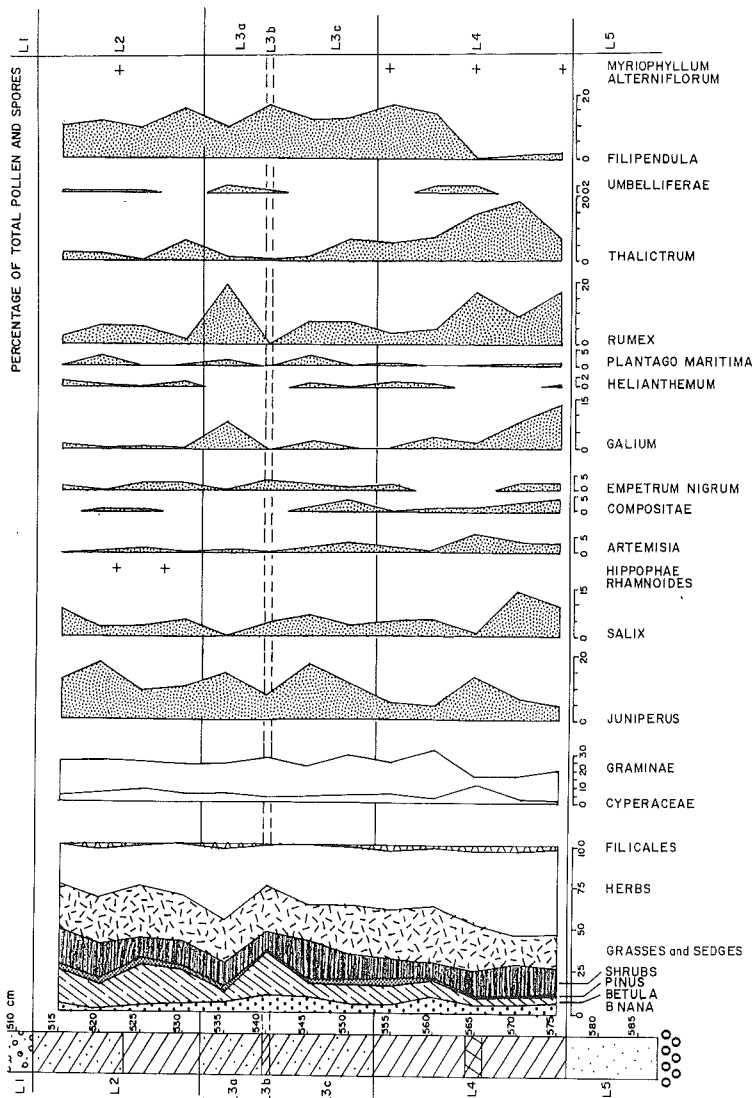


Figure 45. Preliminary pollen diagram for Garths, Sedbergh (from Gunson, 1966).

values (34%) for the whole of the Lateglacial Period (Fig. 45). *Betula cf nana*, *Juniperus* and the herbs all decline during this brief period.

The upper organic mud (L2) has a high tree pollen (up to 30%) and continuing high values of Graminae and *Filipendula*. Although the Lateglacial herbs decline, their continued presence, plus that of *Juniperus*, suggest that the vegetation had still not achieved a closed woodland. The upper sediment (L1) is a coarse pasty clay with much sand and pebbles with solifluction gravels at the edges; difficulties of sampling have not allowed pollen to be collected from this unit.

Significance

Zone L1 is assumed to be the equivalent of the Loch Lomond (Younger Dryas) Stadial with the underlying organic sediments of L2 being ascribed to the Allerød Interstadial (Godwin pollen zones III and II respectively)(Gunson, 1966). The lower part of the sequence is problematical. It could be the result of instability in the steep-sided kettlehole which has caused rapid repetition of the stratigraphy. Whilst this is difficult to completely reject, it is regarded as highly unlikely since the pollen spectra are inconsistent with direct repetition, although a mixture of secondary and contemporary pollen deposition cannot be disproved.

At the time this research was being carried out, the Bölling Interstadial had not been clearly established in Britain, although a number of pollen sites with an earlier climatic oscillation than the Allerød had been identified in the southern Lake District (Smith, 1958), north Lancashire (Oldfield, 1960) and in the Cumberland Lowlands (Walker, 1966). This may be marked by the high arboreal pollen count from Zone L3b, although no radiocarbon dates are available to strengthen this correlation.

Combe Scar

A.R. Gunson & W.A. Mitchell

Introduction

Combe Scar (SD 680878) is situated at the extreme northeast part of the plateau of Middleton Fell which rises to 570m OD on the southern side of Dentdale, west of Barbondale (Fig. 22) and on the western side of the Dent Fault. The cirque is eroded in Coniston grits with an orientation of the cirque axis of 33° . These cliffs are up to 180m in height with numerous gullies and associated debris flows which extend beyond the base of the cliffs. The altitude of the top of the cliffs is 470m OD and the base of the cliff is c. 290m OD. This is an important location since within this cirque there is a very steep, fresh terminal moraine ridge at a very low altitude. Manley (1959) believed that it was occupied by a glacier during the Loch Lomond Stadial and that it was one of the lowest and most southerly of the local ice masses to have come into existence at this time. An infilled tarn within the moraine ridge may give evidence to confirm this hypothesis and was selected for detailed pollen analysis (Gunson, 1966).

Combe Scar moraine ridge

North of the base of the cliff, there is an area of drift which reaches its widest extent to the east of the stream which cuts through the feature to show that it is composed of diamict. Two moraine ridges and a number of mounds may be identified within this area (Fig. 46). The altitude of this moraine is 260m OD. The main ridge is 400m in length and extends from the western side of the cliffs (SD 67758795), eastwards towards Combe House (SD 68158752), where the ridge terminates in an area of mounds. The ridge is asymmetrical in cross profile with a proximal height of 10m with a slope of 10° and a distal slope which rises with a more gentle slope. Near the stream the ridge is more symmetrical but with steeper ($>15^\circ$) slopes. On the western side of the stream, there is a short section of ridge a short distance to the north of the main ridge (Fig. 46).

Combe Scar pollen site

There is a complex sequence of infilled basins within the cirque (Gunson, 1966). In

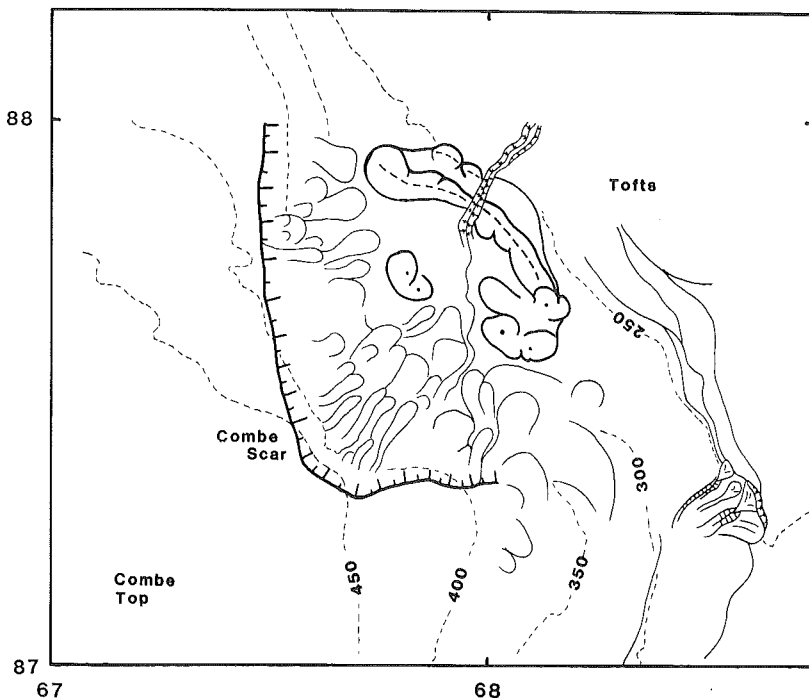


Figure 46. Geomorphological map of Combe Scar (from Mitchell, 1991). Scale and orientation given by National Grid coordinates.

the main basin, lacustrine sediments at the base indicated that the moraine ridge had previously dammed a small lake (Gunson, 1966). A series of borings was taken along the middle of the long axis of this basin (Fig. 47) to establish the deepest part of the basin (Site B).

A typical lithostratigraphic sequence (Core B) (Fig. 47) is as follows:

0-230 cm	carr peat
230-330 cm	coarse detritus mud with <i>Phragmites</i> and <i>Carex</i> fruitstones
330-480 cm	fine detritus mud with <i>Potamogeton</i> fruitstones
480-487 cm	clay-mud transition
487-510 cm	stiff grey clay with much sand
510-520 cm	gravel

The basal deposits were always coarse gravels, above which was a grey clay layer

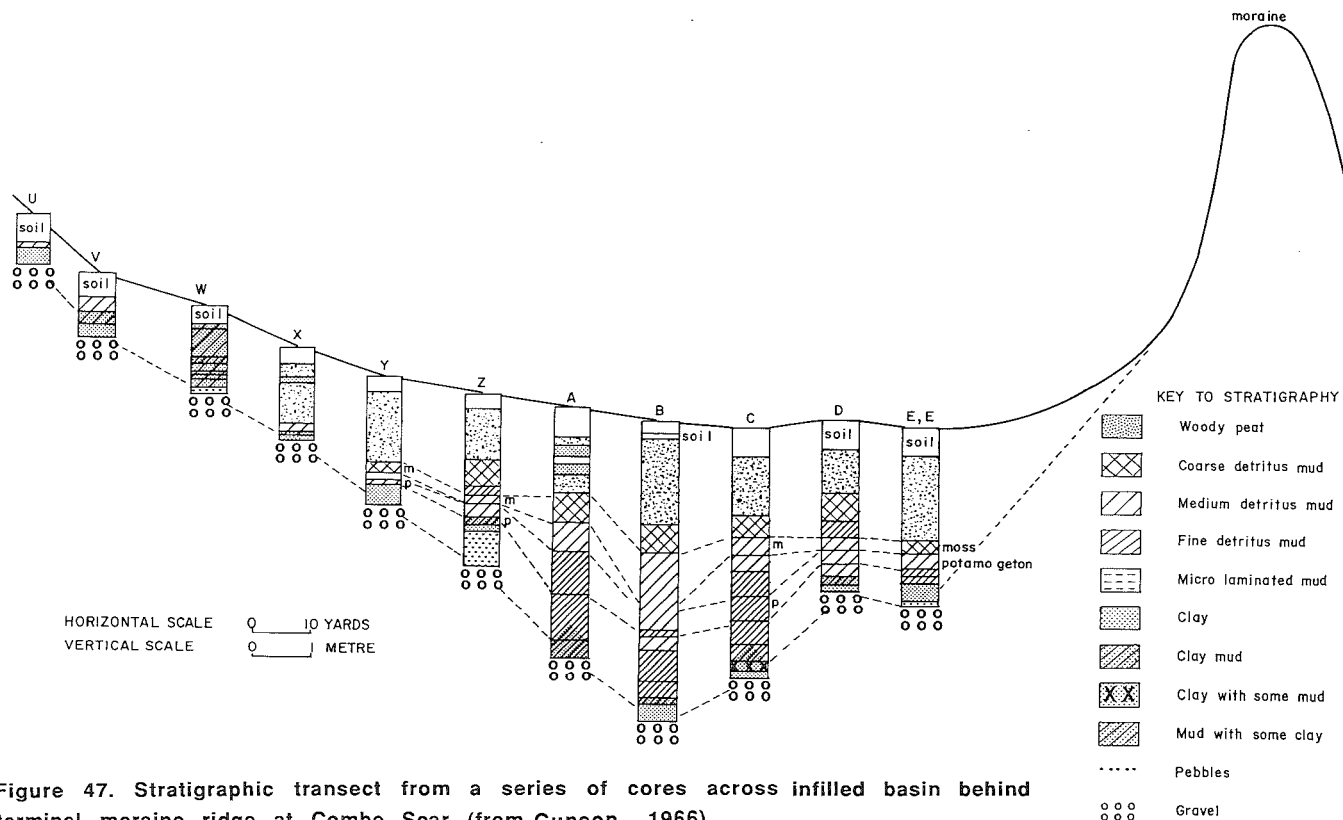


Figure 47. Stratigraphic transect from a series of cores across infilled basin behind terminal moraine ridge at Combe Scar (from Gunson, 1966).

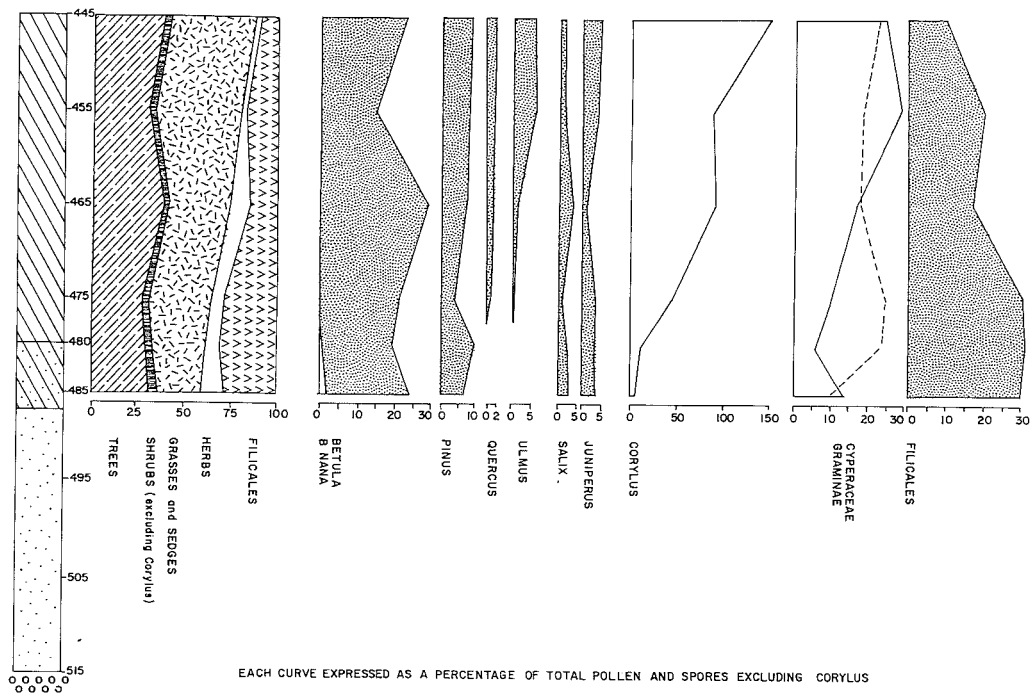


Figure 48. Preliminary pollen diagram for Combe Scar, Dentdale (from Gunson, 1966).

of variable thickness. The clay-mud transition was generally marked by a red-brown nekron mud, although occasionally a thin coarse detritus mud was encountered with *Phragmites* and plant remains (Gunson, 1966).

Pollen Analysis

Samples were taken from the lowest 2m of core from B (Fig. 47). The lowest samples within the clay contained too little pollen to count and the diagram therefore starts at 485m towards the base of the clay-mud transition. Samples were taken at every 5cm in the lowest part of the core up to 475 cm, and then at every 10cm up to the 445 cm level (Fig. 48).

Arboreal pollen is already high in the lowest sample (485 cm) with high values of *Betula* recorded throughout (Fig. 48). Towards the top (465 cm), *Ulmus* is present in medium quantities and *Quercus* is continuous. The greatest individual change is shown by *Corylus* which is present at the base and rises to 150% total arboreal pollen at 475 cm and to 400% at 445 cm (Fig. 48). Graminae and Cyperaceae show high values throughout the core and rise towards 445 cm while Filicales are high at the base and show a gradual decline up the sequence.

The high *Betula* counts and the low herbaceous values in the basal counts indicate that the base of the diagram is already well into a Postglacial sequence, probably Godwin zone IV-V of the British sequence, while the high *Corylus*, continuous *Quercus* curve and *Ulmus* rise suggest that the top of the diagram is within Zone VI (Fig. 48).

Significance

The pollen analysis and stratigraphical evidence is consistent with a small glacier having occupied the cirque during pollen Zone III (Loch Lomond Stadial) confirming Manley's proposal of a former ice mass in this cirque at this time.

Palaeoclimatic Reconstruction

The former glacier which occupied Combe Scar during the Loch Lomond Stadial produced a large arcuate moraine ridge at a very low altitude (260 m). The glacier has been reconstructed from the geomorphic evidence (Fig. 49) and shows that it covered 0.22 sq. km, and achieved a maximum thickness of 60m (Table 3). Glacial

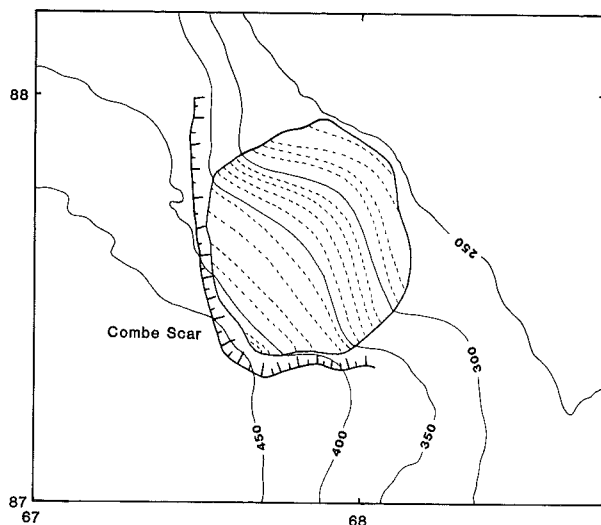


Figure 49. Reconstruction of the Combe Scar palaeoglacier for 'normal' mode (in mass balance equilibrium); former surface contoured at 10m intervals (from Mitchell, 1991).

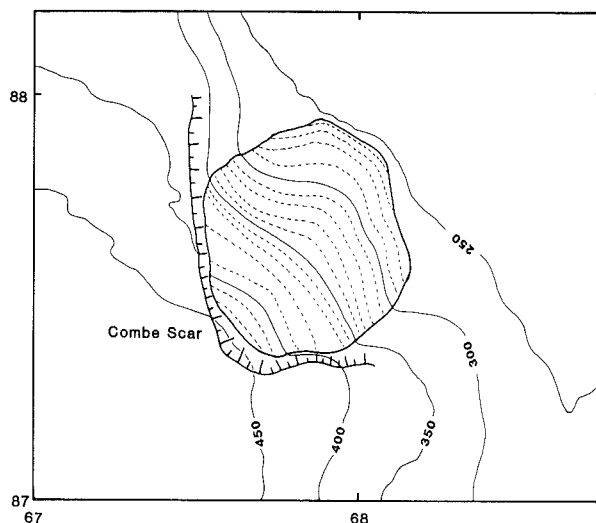


Figure 50. Reconstruction of the Combe Scar palaeoglacier for 'advance' mode (thicker glacier towards snout); former surface contoured at 10m intervals (from Mitchell, 1991).

reconstruction of the palaeoglacier gives an ELA value of 311m (Table 5). The very low ELA value is explained by the high potential snowblow areas which lie to the southwest of the cirque (Fig. 24) and which gives this glacier the highest mean snowblow factor of 7.95 and with a maximum value of 6.91 in the southwest (181-270°)(Table 6). The ratio of potential snowblow area to actual glacier size is very high (63.3)(Table 6).

These values clearly show that both the western glaciers of Cautley Crag and Combe Scar were controlled in their development and sustenance by snow accumulation associated with large plateau areas upwind from the glacier locations. In the case of Combe Scar this is clearly associated with southwesterly winds.

Palaeoglaciological Reconstruction

This palaeoglacier gives some of the highest results with maximum ice thicknesses at the ELA of 44m which with the steep slope of the glacier gives a surface velocity of 3.46ma^{-1} (Table 7). Reconstructing this glacier in 'advance mode' (Fig. 50) gives a reasonable increase in surface velocity to 8.76ma^{-1} which makes it the fastest of the five reconstructed glaciers. If this is taken to consider mass flux through the ELA cross section, then the glacier must have had a high mass balance to maintain this velocity and have an ELA at a low altitude; this suggests that high amounts of snow would have been required to have blown off the plateau snow field to maintain this particular glacier (Mitchell, 1991).

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Quaternary Research Association

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The main meetings of the Association are the Annual Field Meeting, usually lasting 3 or 4 days, held in April, and a 1 or 2 day Discussion Meeting held at the beginning of January. Additionally, Short Field Meetings may be held in May or September and occasionally these visit overseas locations. Short Study Courses on the techniques used in Quaternary work are also held. The publications of the Association are the *Quaternary Newsletter* issued with the Association's *Circular* in February, June and November; the *Journal of Quaternary Science* published in Association with Wiley, and with four issues a year; the Field Guides Series and the Technical Guides Series.

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